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Water Resources Center

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ABSTRACT

Increasing demands for water supply have accompanied rapid population growth in the Las Vegas Valley and portions of surrounding southern Nevada. Exploration and development of groundwater resources to meet these demands increases the potential for impact on groundwater systems to the north and west of the Lake Mead National Recreation Area. Because the park is located down-hydraulic-gradient from these areas, large-scale changes in groundwater use may affect groundwater resources and, ultimately, discharge from natural springs within the park. This study was conducted for the National Park Service to investigate the hydrology and hydrogeochemistry of selected springs in the Lake Mead and Black Canyon areas, and to determine the source areas associated with these springs.

Thirty six springs were visited and described. Historic geochemical data were compiled and supplemented by new stable and radioactive isotopic data. Three classifications of source area were defined, primarily based on hydrogeologic setting and stable isotopic data. Almost one third of the springs were found to discharge from local groundwater systems, many of which are entirely contained within the park boundaries. These springs are generally not related to major structural features and their stable isotopic values indicate that they receive most or all of their recharge locally and at low elevations, despite the minimal groundwater recharge generally assumed for low elevations in southern Nevada. A second set of springs was found to discharge groundwater that originates outside local flow systems, and therefore outside the park boundaries. Many of these springs are related to major, regional structural features, and their stable isotopic values are indicative of recharge at elevations higher than most of the region surrounding Lake Mead, although they do not appear to be directly related to regional groundwater flow from the White River Flow System or the Virgin River basin. Data obtained from a third set of springs, located below Hoover Dam in Black Canyon, suggests that these springs are strongly influenced by recirculated Lake Mead water, confirming earlier work.

CONTENTS

TABLES

FIGURES

INTRODUCTION

Springs on the western edge of Lake Mead and in the Black Canyon of the Colorado River are important natural hydrologic features of the Lake Mead National Recreation Area. Although many springs are little more than seeps, their discharge represents the only available perennial surface flow in large portions of this arid region. These springs appear to originate from a variety of sources ranging from precipitation in local drainage basins to regional interbasin groundwater flow systems.

Rapid population growth in portions of southern Nevada, particularly in the Las Vegas Valley, has increased the need for additional water supplies in the area, including groundwater. As a result, there has been a dramatic increase in the potential for additional large-scale development of groundwater resources to the west and north of Lake Mead, areas which are hydraulically upgradient of many of the springs. If large-scale development of groundwater resources occurs in source areas or along flow paths leading to springs, the discharge of these springs could be impacted.

To address concerns regarding potential impacts on spring resources, and to plan for their management and protection, the National Park Service (NPS) requires scientific information on the hydrology and hydrogeochemistry of springs near Lake Mead, and particularly whether the waters are of local or regional origin. This investigation was undertaken to: 1) provide a comprehensive database of spring chemical and isotopic composition; and 2) determine the source areas of and flow paths to selected springs.

Geography and Climate

The waters of the Colorado River impounded by Hoover Dam form Lake Mead and divide southeastern Nevada from northwestern Arizona (Figure 1). The lake is located near the transition between the Great Basin and Colorado Plateau physiographic provinces. Elevations in the region adjacent to the lake are generally less than 1000 m (all elevations given in this report are referenced to mean sea level), and range from about 200 m at the Colorado River below Hoover Dam to over 1600 m in the Muddy Mountains. The highest mountain ranges in southern Nevada are the Spring Mountains (3630 m) and the Sheep Range (3020 m) which rise 60 km to the west and northwest, respectively, of Lake Mead.

The climate is one of extremes, ranging from arid in the low elevation basins, where the highest temperatures and lowest precipitation amounts in the Great Basin occur, to sub-humid in the higher mountains. In the Las Vegas Valley, the mean summer temperature at an elevation of 640 m is 30.8°C and the mean annual precipitation is 10.4 cm (Western Regional Climate Center, 1997). Orographic effects cause precipitation amounts to increase with elevation such that the upper elevations of the Spring Mountains receive up to 70 cm of precipitation annually (Malmberg, 1961).

Annual precipitation trends show a pronounced seasonality, with maximum amounts typically received in December and August. Winter precipitation generally falls as long-duration, low-intensity frontal storms derived from moisture moving eastward from the Pacific Ocean, while summer precipitation originates to the south in the Gulf of California and the Gulf of Mexico and is often delivered as short-duration, intense thunderstorms (Quiring, 1965; French, 1983). The rainshadow effect of the Sierra Nevada Mountains in the winter and the incomplete flow of moisture

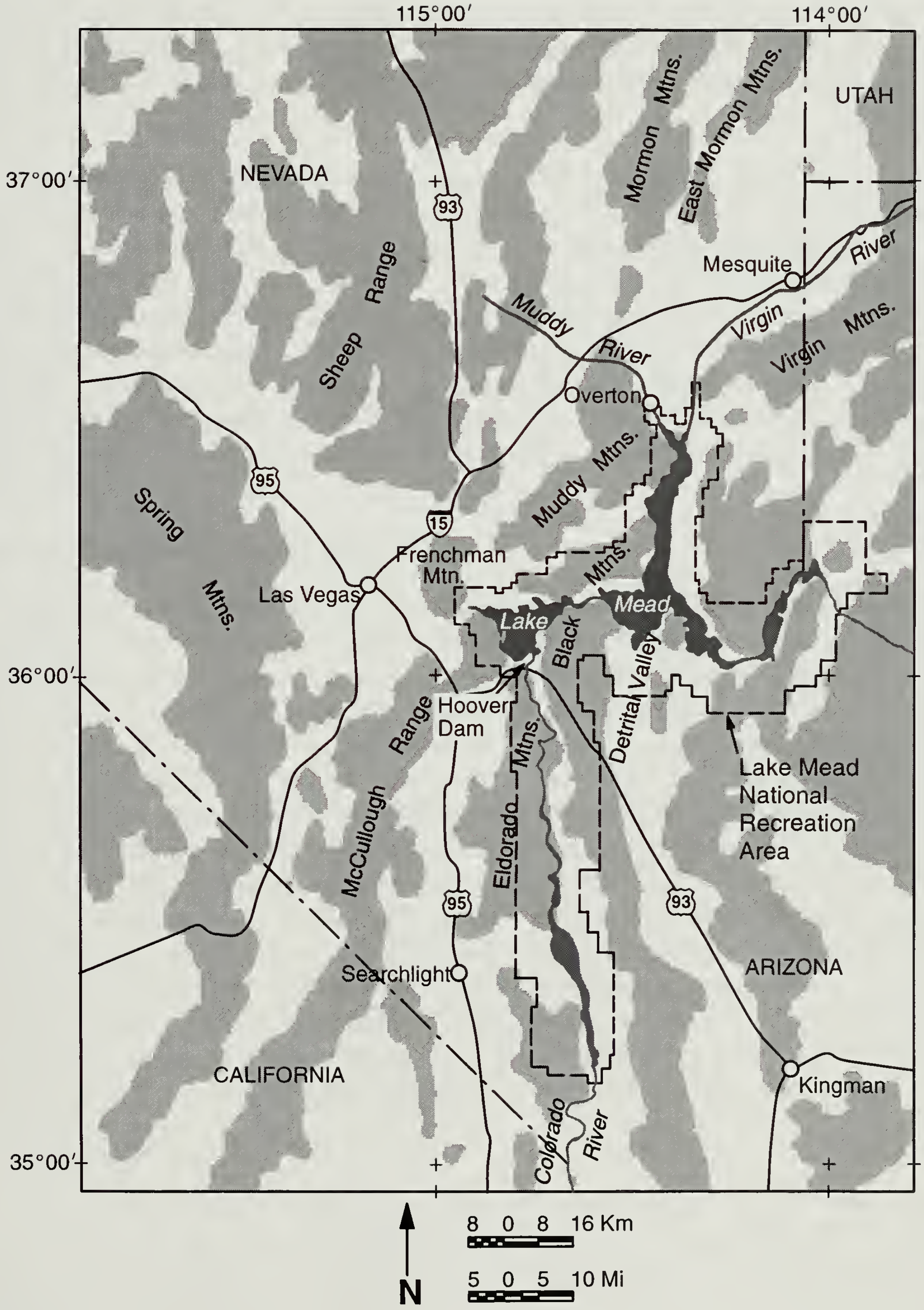

Figure 1. Location of the study area in southeastern Nevada and northwestern Arizona.

from the south in the summer forms a zone of precipitation deficit in the western portion of southern Nevada (Quiring, 1965). The eastern portion is less affected by the Sierra Nevada rainshadow and is open to the flow of moisture from the south in the summer, thus causing a zone of precipitation excess.

Estimates of groundwater recharge from precipitation in Nevada are commonly developed using the Maxey-Eakin method (Maxey and Robinson, 1947; Maxey and Eakin, 1949), which is based on empirically-derived relationships between precipitation and recharge in several groundwater basins in the state. In the Las Vegas Valley, the Maxey-Eakin method predicts that groundwater recharge is negligible where annual precipitation is less than 25.4 cm, corresponding to elevations below approximately 1800 m (Maxey and Robinson, 1947). Below this elevation, the estimated annual precipitation volume is calculated to be lost to evapotranspiration (due to high air temperatures and low humidity) and surface runoff (due to sparse vegetation and low-permeability soils). Thus, on the scale of groundwater basins, recharge is considered to be minimal in much of southern Nevada.

Previous Studies of Springs in the Region

Chemical and isotopic data are available for numerous springs in southeastern Nevada, primarily as a result of the Nevada Carbonate Aquifer Program studies. Lyles *et al.* (1987) compiled chemistry data for wells and springs in Nevada within a 160 km radius of Las Vegas. Thomas *et al.* (1991) compile a similar database, but include isotopic data collected from wells, springs, and streams. Thomas *et al.* (1997) supplement the earlier database with data from additional sampling sites, describe chemical and isotopic processes and composition of groundwater in basin-fill and carbonate aquifers, and delineate flow systems in the carbonate rocks of southern Nevada. Studies of hydrogeologic resources pertinent to the present study have been conducted by Laney (1981) and Laney and Bales (1996) as part of an ongoing series of reconnaissance studies of the Lake Mead National Recreation Area. These reports provide physical descriptions, geologic setting, and chemical data for many of the springs. In the only detailed interpretive study of springs within the recreation area, McKay and Zimmerman (1983) investigated springs in Black Canyon using hydrogeochemical, stable isotope, and tritium data. Finally, the Southern Nevada Water Authority (SNWA) has initiated an investigation of the origins of groundwater issuing from springs on the Nevada side of Black Canyon, collecting extensive chemical and isotopic data.

Acknowledgements

The authors wish to thank Paul Christensen of the National Park Service, Water Resources Division, for facilitating this research and providing guidance in the early going. Bill Burke and the Resource Management staff at the Lake Mead National Recreation Area are thanked for providing background information on the history and locations of the springs. Alan McKay of the Desert Research Institute and James Thomas of the U.S. Geological Survey are thanked for invaluable insights into the hydrogeology of the region. The majority of this work was funded by the National Park Service.

METHODOLOGY

The chemistry of groundwater is a result of the type and amount of minerals present in the rocks through which the groundwater moves, and the conditions of recharge and discharge. Generally, groundwater chemistry evolves along flow paths from recharge areas to discharge areas as geochemical reactions occur between the water and rock. At the local scale, however, local geologic complexity can lead to large variations in groundwater chemistry.

Although flow paths that supply groundwater to springs can be described using the geochemistry of spring discharge, delineation of recharge sources is often more effectively approached using the spring's isotopic composition. Because the principal objective was to delineate groundwater source areas, this study focused on several stable and radioactive isotopes in groundwater. Ratios of the stable isotopes in water molecules, oxygen-18 (^{18}O) to oxygen-16 (^{16}O) and deuterium (D) to hydrogen (^{1}H), often provide more definitive identification of source areas for groundwater than water chemistry. In addition, the radioactive isotopes tritium (^{3}H) and carbon-14 (^{14}C) can be used to determine relative ages of groundwater. A relatively young age reflects the dominance of local recharge and short residence times, while an older age reflects a longer residence time and often indicates lengthy travel times in regional flow systems. Finally, radioactive isotopes of uranium (^{234}U and ^{238}U) can be used for tracing groundwater masses from recharge areas to discharge areas. Background information on these techniques is provided below for ease of reference.

The stable isotopes D and ^{18}O are useful tools for tracing groundwater because, unlike major ion geochemistry, stable isotopic composition is essentially unchanged by the rocks through which groundwater travels (under non-geothermal conditions). The stable isotopic composition of groundwater recharge is related to the temperature, amount, distance from the ocean, and altitude of precipitation (Mazor, 1997), therefore, groundwaters originating in a common source area often share similar stable isotopic composition. Stable isotopes are particularly useful in this study because the pervasive gypsum deposits and other evaporites in the region cause dramatic changes in groundwater geochemistry near spring discharge areas, effectively masking the original geochemical composition of the groundwater.

The stable isotope ratio $^{13}C/^{12}C$ (expressed in a delta notation as $\delta^{13}C$) is very sensitive to biologic processes and thus there can be large differences in $\delta^{13}C$ of carbon subjected to differing photosynthetic, bacterial and other processes. Recharge water, percolating through soils, dissolves CO_2 gas that has a $\delta^{13}C$ signature characteristic of the local plant cover. Reactions with carbonate rocks impart enriched $\delta^{13}C$ values, sensitive to the carbonate origin in pedogenic and marine deposits. In addition to this tracing function of $\delta^{13}C$, the isotope is also used to correct ^{14}C groundwater ages for dilution by dissolved rock carbon. ^{14}C is a radioactive isotope present in dissolved inorganic carbon in groundwater. As such, ^{14}C does not provide a direct age measurement of the water, as tritium does, but requires an understanding of the source of the dissolved inorganic carbon for correct interpretation (Mook, 1980). The long half-life of ^{14}C (5730 years) makes it useful for dating groundwaters with residence times in excess of several decades.

The radioactive isotope tritium provides a semi-quantitative means for dating groundwater with residence times of several decades or less (Mazor, 1997). Groundwaters having tritium concentrations below 5 pCi/L are considered to be derived primarily from recharge prior to the onset of atmospheric testing of nuclear bombs in 1952, while groundwaters having concentrations greater than 5 pCi/L are considered to have at least some component recharged after 1952. Due to its short half life (12.3 years), tritium concentrations in atmospheric precipitation have declined since the period of maximum testing in 1962. In 1994 through 1996, tritium concentrations in southern Nevada precipitation ranged between 10 and 20 pCi/L in the winter and between 20 to 60 pCi/L in the summer (Dennis Farmer, U.S. EPA, personal communication). This cycle between winter lows and summer highs is observed worldwide and is related to the circulation of moisture in the upper atmosphere (Roether, 1967).

The radioactive isotopes of uranium can be useful groundwater tracers because of their high solubility, insensitivity to chemical reactions, and long half-lives (Osmond and Cowart, 1976; Cowart, 1979). They are especially useful in southern Nevada because of the wide range of natural uranium concentrations in the groundwaters of the region (Farmer, 1996). Since uranium is presently less widely-used for tracing groundwater than the isotopes described above, a more detailed description of the method follows. Uranium is a naturally-occurring element which dissolves in groundwater when dilute recharge waters interact with uranium-bearing minerals in the subsurface. The vast majority (99.725 percent) of natural uranium occurs as the isotope ^{238}U, which has a half-life of 4.46×10^9 years. The radioactive decay of ^{238}U produces ^{234}U, which comprises about 0.005 percent of naturally-occurring uranium, and has a half-life of 2.45×10^5 years.

The activity of a radionuclide is defined by the equation $A = N\lambda$, where A is the activity of any radionuclide, N is the number of atoms of that nuclide present in the system being examined, and λ is the decay constant for that nuclide (Osmond and Cowart, 1976). The value of λ indicates the number of disintegrations an isotope undergoes per unit time, and is thus inversely proportional to the half-life of an isotope. The activity equation shows that two radionuclides that have significantly different numbers of atoms present in a system can have the same activities if their half-lives are sufficiently different. This proves to be the case with ^{234}U and ^{238}U, which, in closed geologic systems (such as unweathered rocks), tend to achieve a state known as secular equilibrium, where the activity of ^{234}U (low number of atoms, but relatively short half-life causing a high number of decays per unit time) and that of ^{238}U (high number of atoms, but relatively long half-life causing a low number of decays per unit time) become equal. It takes approximately 10^6 years from the time of formation for a system to achieve this secular equilibrium (Osmond *et al.*, 1968).

^{234}U and ^{238}U tend to achieve secular equilibrium in closed geologic systems. However, in natural rock-groundwater systems, disequilibrium between ^{234}U and ^{238}U is quite common (Thurber, 1962) and thought to be present due to side effects resulting from the radioactive decay process (Gascoyne, 1992). Disequilibrium is typically quantified via the $^{234}U/^{238}U$ activity ratio (AR). A system in secular equilibrium would have an AR equal to one; a system with "excess" ^{234}U activity would have AR greater than one, and a system with "excess" ^{238}U would have an AR less

than one. The majority of groundwaters exhibiting disequilibrium show AR greater than one, indicating an excess of ^{234}U (Osmond and Cowart, 1976).

Uranium has two naturally occurring valence states (+4 and +6). U^{6+}, which is present in oxidizing conditions, is soluble, while U^{4+}, which predominates in reducing conditions, has an extremely low solubility, and is thus considered immobile. The presence of reducing conditions can greatly complicate the analysis of uranium, but the waters sampled for this study consistently showed dissolved oxygen content indicative of oxic waters (Table A-1). Although deep groundwater is typically thought to be anoxic, deep waters in Nevada and other parts of the Basin and Range physiographic province are commonly found to be oxic (Winograd and Robertson, 1982).

Most of the springs in the present investigation have been visited and described during the studies described above, and discharge measurements, chemical indicator measurements, and water chemistry analyses are available. However, few of the springs have been sampled for stable and radioactive isotope analysis. The historic inventories and previous studies provided a basis for identifying the locations of springs and for the building of the present database of physical, chemical, and isotopic data. Data collection for the present study focused on isotopic constituents.

All of the springs were visited at least once during the course of this study. Spring coordinates were determined using a Magellan 9500 Pro hand-held GPS unit in autonomous mode. Low discharge rates were measured using a beaker and stopwatch and high discharge rates were measured using a Marsh-McBirney Flo-Mate 2000 flow meter. Field measurements were made of temperature, pH, electrical conductivity (EC), dissolved oxygen (DO), and alkalinity (HCO_3) using standard field analytical equipment. The physical, chemical, and isotopic data derived from previous studies, and data collected for the present study, are compiled in Appendix A. Geologic descriptions and sketch maps were developed for each spring area and are included in Appendix B. Isotopic data for selected southern Nevada groundwaters are compiled in Appendix C.

This report describes thirty-six springs which are located in two general areas (Table 1). One is the Lake Mead basin, including the area west of the Overton Arm and the area north of Lake Mead (Figure 2). The other is the area of the Black Canyon of the Colorado River, downstream of Hoover Dam (Figure 3).

GEOLOGIC SETTING

The Lake Mead National Recreation Area is located near the eastern margin of the Basin and Range geologic province, a region comprised of broad, flat-lying valleys underlain by thick alluvial deposits and bordered by narrow, nearly parallel mountain ranges. Situated between mountain ranges composed of Paleozoic to Mesozoic sedimentary rocks and a Precambrian terrain intruded by Cenozoic igneous rocks (Figure 4), the recreation area lies near the southeastern end of the regional carbonate-rock aquifer system. This large aquifer system is defined as the area where 80 percent of the measured section is over 50 percent carbonate rock (Mifflin, 1968), and underlies 260,000 km^2 of eastern Nevada, western Utah, southeastern Idaho, and extreme southeastern California (Dettinger, 1989). Table 2 presents a simplified stratigraphic column used in the present study.

Table 1. Identification Numbers and Names of Springs Included in this Study. Names in the Lake Mead basin are official names. Names in Black Canyon are unofficial names given by McKay and Zimmerman (1983), with the exception of springs given unofficial names by the National Park Service.

ID	Name	Comments
		Lake Mead Basin
1	Kelsey Spring	
2	Unnamed	Located in Magnesite Wash
3	Unnamed	Located in Kaolin Wash
4	Getchel Spring	
5	Unnamed	Uppermost Spring in Valley of Fire Wash
6	Unnamed	Upper Spring in Valley of Fire Wash
7	Unnamed	Lower Spring in Valley of Fire Wash
8	Blue Point Spring	
9	Unnamed	Located 0.8 km south of Spring 8
10	Unnamed	Located 0.8 km southeast of Spring 9
11	Rogers Spring	
12	Scirpus Spring	
13	Corral Spring	
14	Unnamed	Located northwest of Rogers Bay
15	Bitter Spring	
16	Sandstone Spring	
17	Cottonwood Spring	
18	Gypsum Spring	
19	Unnamed	South of Rainbow Gardens
		Black Canyon
20	Pupfish Spring	
21	Arizona Hot Spot	
22	Sauna Cave	
23	Nevada Hot Spring	NPS name, "Fort Lucinda" of McKay and Zimmerman (1983)
24	Nevada Hot Spot	
25	Palm Tree, Hot	
26	Palm Tree, Cold	
27	Unnamed Spring	Located in Horsethief Canyon
28	Boy Scout Canyon, Hot Spring	NPS name, "Rifle Range" of McKay and Zimmerman (1983)
29	Boy Scout Canyon, Cold Spring	
30	Arizona Hot Spring	NPS name, "Ringbolt Rapids" of McKay and Zimmerman (1983)
31	Unnamed	Cold Spring located near Arizona Hot Spring
32	Nevada Falls	
33	Bighorn Sheep Spring	
34	Arizona Seep	
35	Latos Pool	
36	Unnamed	Located in Aztec Wash

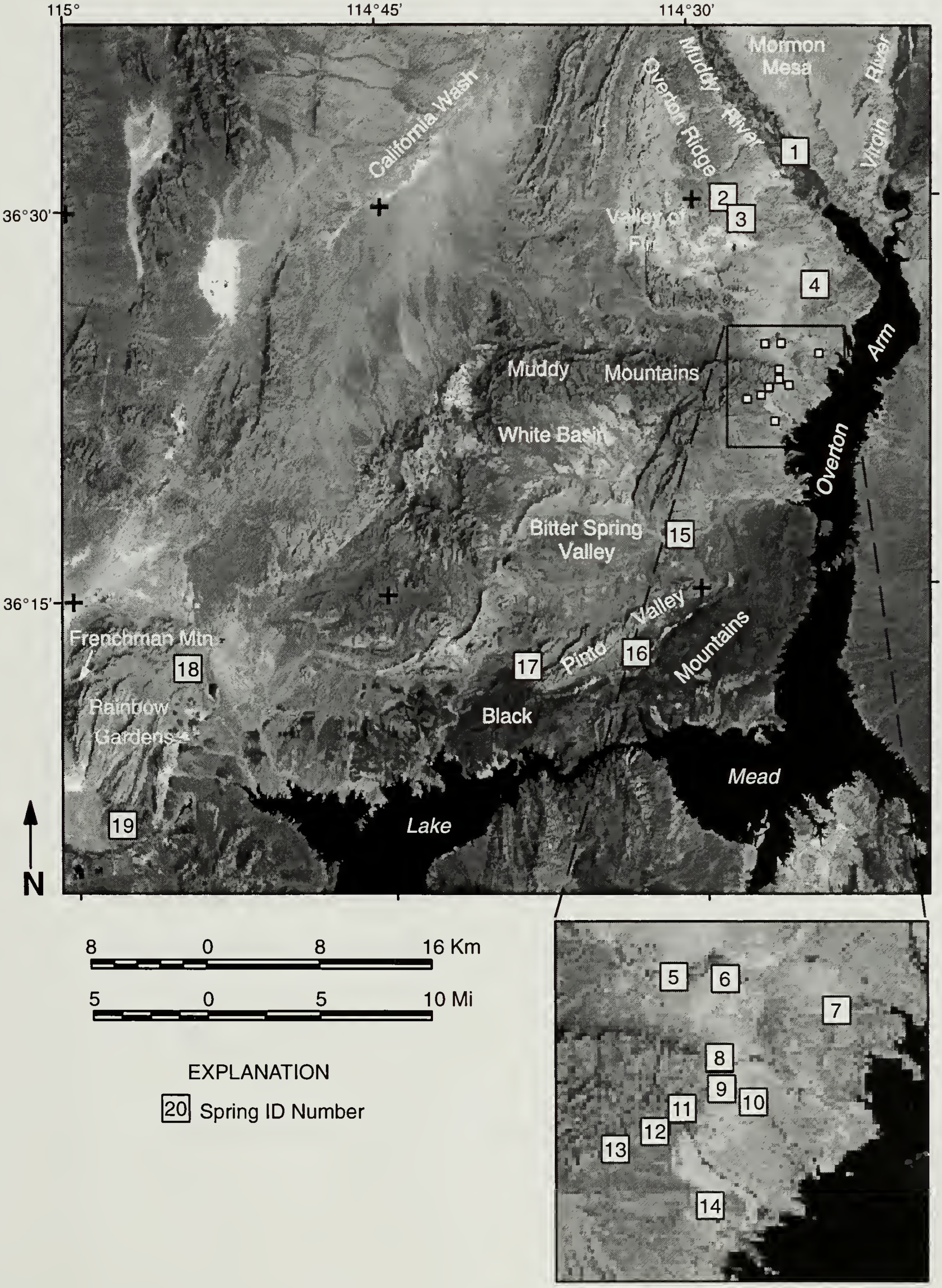

Figure 2. Locations of springs in the Lake Mead basin. Detail shows springs in the North Shore complex.

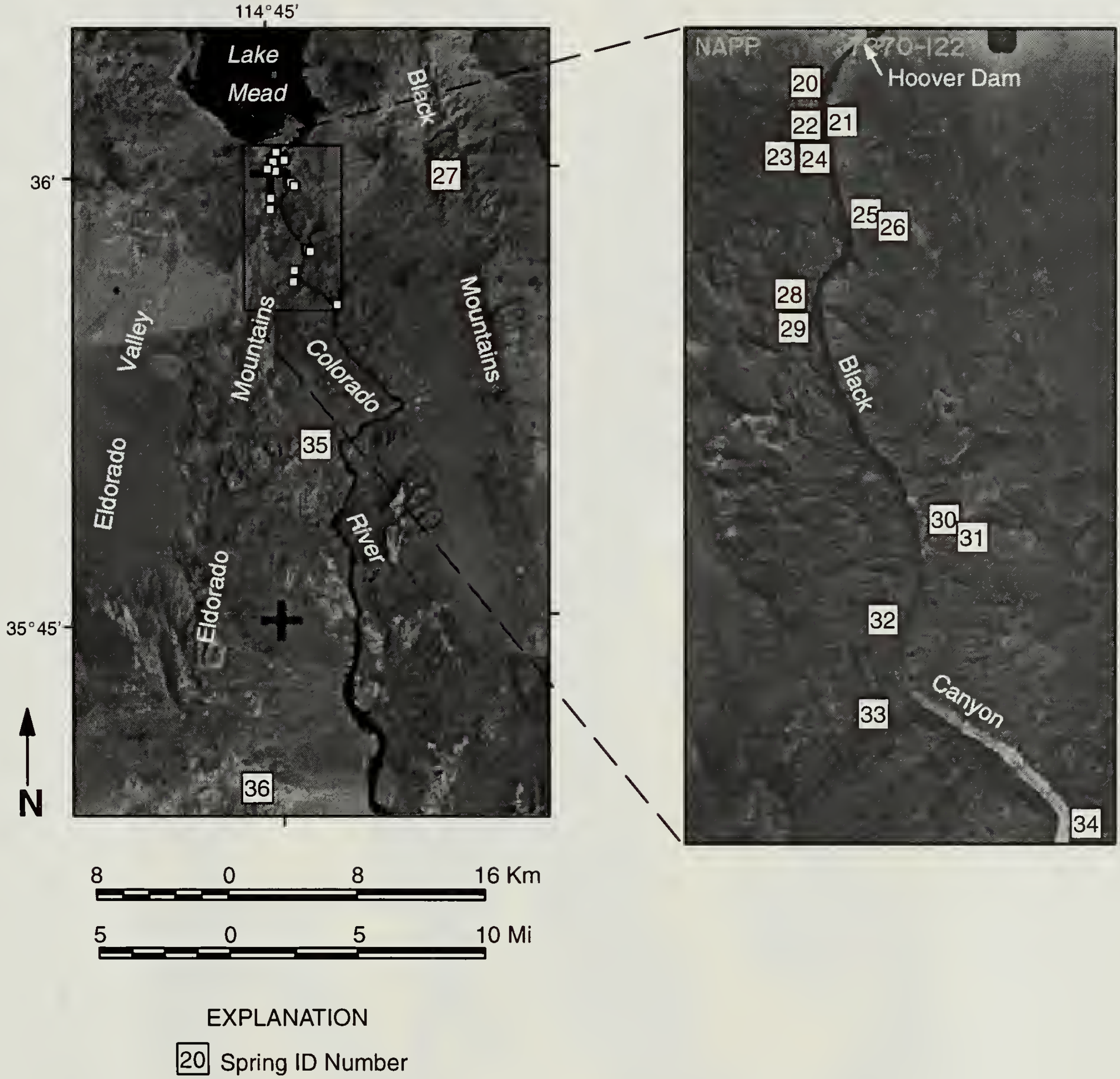

Figure 3. Locations of springs in the Black Canyon area. Detail shows springs in Black Canyon proper.

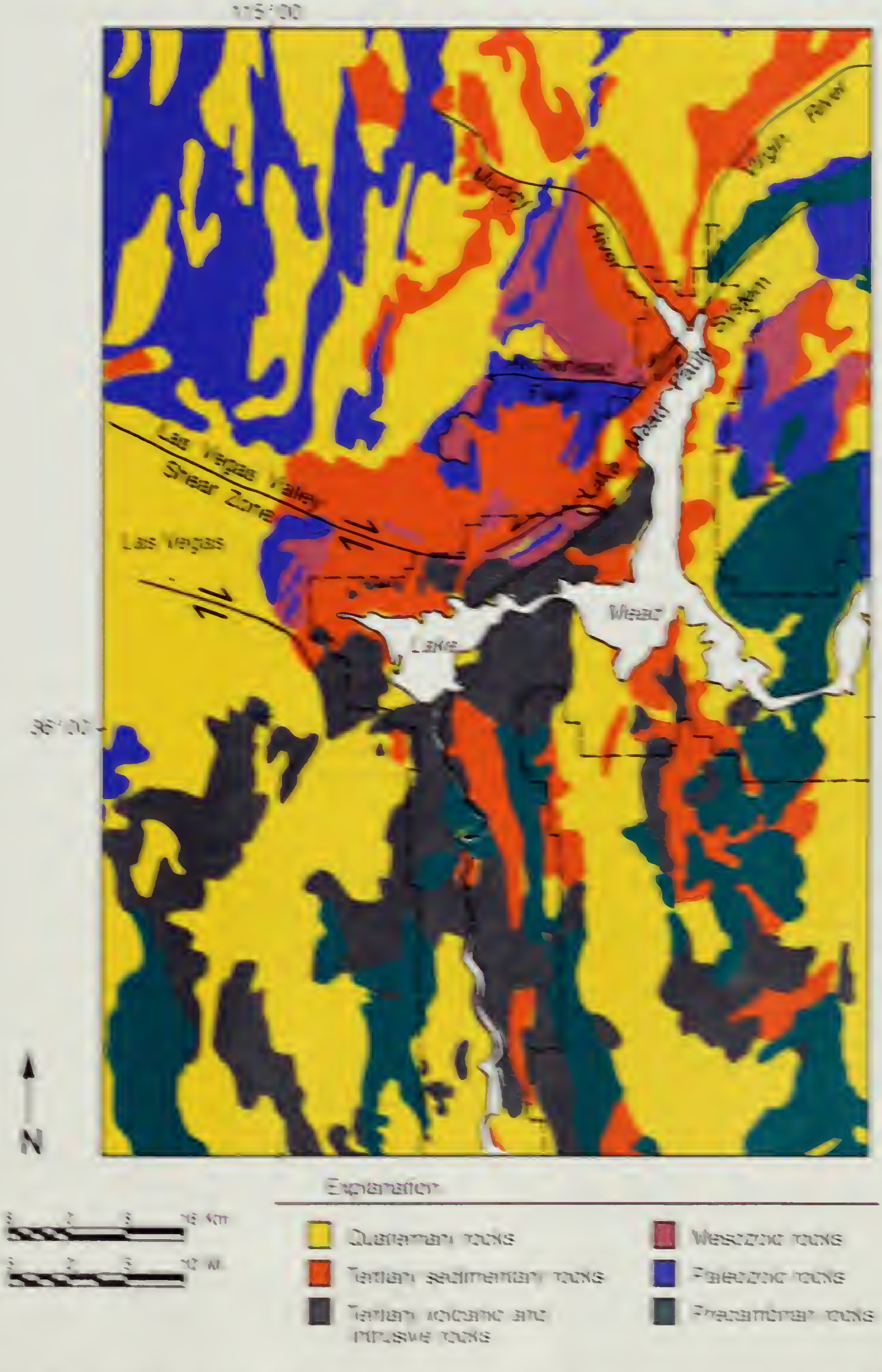

Figure 4. Generalized geologic map of southeastern Nevada and northwestern Arizona. Modified from Longwell et al. (1965), Reynolds (1988), and Campagna and Aydin (1994).

Table 2. Generalized Stratigraphic Column for the Study Area.

Time		Unit	Symbol	Description and Reference
Cenozoic	Quaternary	Alluvium (Holocene to Pleistoncene?)	Qal	Silts, sands, pebbles, cobbles, and boulders in modern drainages. Angular to subrounded particles. Unconsolidated, locally derived. (Bohannon, 1984).
		Older Alluvium (Pleistocene)	Qoa	Silt, sand, pebbles, cobbles, and boulders in alluvial fans, thick colluvial deposits, alluvial flood plains, and channels. Poorly sorted, angular to subround unconsolidated particles. Locally derived. (Bohannon, 1984).
		Terrace Deposits (Pleistocene?)	Qt	Silt, sand, pebbles, cobbles, and boulders. Compacted and/or cemented. Locally derived. (Bohannon, 1984).
	Tertiary	Miocene Volcanics (undifferentiated)	Tmv	Lava flows of Callville Mesa and Overton Arm and intrusive rocks north of Callville Mesa, western Bitter Spring Valley, and northeastern Muddy Mountains. (Bohannon, 1984).
		Muddy Creek Formation	Tm	Bedded siltstone, sandstone, gypsum, gypsiferous siltstone, and conglomerate near basin margins. (Bohannon, 1984).
		Horse Spring Formation	Th	Limestone, dolomite, conglomerate, sandstone, volcanic tuff, gypsum and breccia. Includes clastic and gypsum facies of the Thumb Member. (Bohannon, 1984).
		Rainbow Gardens Basal Conglomerate	Thrc	Conglomerate consisting of sandstone, siltstone, gypsum, gypsiferous siltstone, carbonates, and magnesite. Lowest unit in the Horse Spring Fm, and marks the Tertiary unconformity. (Bohannon, 1984).
		Mount Davis Volcanics (undifferentiated)	Td	Miocene lava and flow breccias. (Anderson, 1978).
		Intrusive Rocks (undifferentiated)	Ti	Miocene intrusive rocks. Includes the Boulder City pluton, a mixture of medium-grained granodiorite and andesitic border facies (Anderson, 1969), and the Wilson Ridge pluton, a biotite granite through horneblende-biotite granodiorite to pyroxene-biotite diorite. (Anderson, 1978).
		Patsy Mine Volcanics (undifferentiated)	Tpv	Miocene. In the study areas andesitic lava and breccia. (Anderson, 1978).
Mesozoic	Jurassic-Cretaceous	Autochthonous Jurassic and Cretaceous Formations	JKau	Baseline Sandstone (K): sandstone and conglomerate. Willow Tank Formation (K): Conglomerate, claystone, sandstone, tuff, and mudstone. Aztec sandstone (J, K?): red quartz arenite w/hematite cement. (Bohannon, 1984).
	Triassic	Autochthonous Triassic Formations	₮au	Moenave and Kayenta Formations: gypsiferous sandstone and siltstone. Chinle Formation: sandstone, siltstone, claystone, conglomerate, minor limestone. Moenkopi Formation: siltstone, sandstone, gypsum, gypsiferous siltstone, limestone, conglomerate. (Bohannon, 1984).

Table 2. Generalized Stratigraphic Column for the Study Area (Continued).

Paleozoic	Permian	Autochthonous Permian Red Beds and Kaibab-Toroweap Formations	Pau	Permian Red Beds (lower P): sandstone, siltstone, gypsum. Kaibab-Toroweap Fms (P): limestone, chert, siltstone, gypsum. (Bohannon, 1984).
	Cambrian-Pennsylvanian	Allochthonous Paleozoic Rocks (undifferentiated)	O|Pal	Bonanza King Fm. (Є) through Bird Spring Fm (P |P): limestone, dolomite, sandstone, quartzite, shale. (After Bohannon, 1984).
Proterozoic	Precambrian	Variegated Metamorphic Rocks	pЄ	Predominantly biotite-almandine gneiss and schist and garnetiferous granite pegmatite. (Anderson, 1978).

The Precambrian/Cenozoic terrain in the southern portion of the study area includes the Black Mountains, the Eldorado Mountains, and Black Canyon. The Precambrian section is comprised of variegated metamorphic rocks consisting of biotite-almandine gneiss and schist and garnetiferous granite pegmatite (Anderson, 1978). These rocks are exposed in the Lake Mead area where structural highs formed during the late Cretaceous to early Tertiary Sevier orogeny resulted in erosion of the overlying Paleozoic and Mesozoic sedimentary rocks (Bohannon, 1984). Tertiary volcanic and intrusive rocks (described below) extensively intrude the Precambrian rocks.

Paleozoic rocks are exposed in the northern portion of the study area in the Muddy Mountains, North Muddy Mountains, and the western portion of Frenchman Mountain. The Paleozoic rocks are predominantly limestone and dolomite (carbonate rocks), with lesser amounts of sandstone, quartzite, and shale. To the northwest, the Paleozoic section reaches a thickness of 5,000 m near the Sheep Range (Longwell *et al.*, 1965) and 7600 m near the Nevada Test Site (Tschanz and Pampeyan, 1970). However, the section thins dramatically eastward in the area west of the Overton Arm, reflecting a hinge line between deep-water and shelf deposits (Stewart, 1970). At the Muddy Mountains, the Paleozoic section is reduced to a thickness of 1200 m (Longwell *et al.*, 1965).

Mesozoic rocks are exposed in the Valley of Fire area, the northern edge of the Black Mountains bordering Pinto Valley, and the eastern portion of Frenchman Mountain. Mesozoic rocks are predominantly sandstones, siltstones, and conglomerates, with varying amounts of gypsum. The Formations exposed in the study area are shown in the stratigraphic column (Table 2).

Tertiary volcanic and intrusive rocks are found within the Precambrian terrain in the southern portion of the study area. The oldest Tertiary rocks are andesitic lava and breccia of the Miocene Patsy Mine volcanic rocks (Anderson, 1971) and are well exposed along the cliffs of Black Canyon. The intrusive rocks include the Miocene-aged Hoover Dam and Wilson Ridge plutons, and numerous dikes of rhyolitic to basaltic composition (Anderson, 1978).

Tertiary sedimentary rocks are exposed throughout the study area, yet predominate in the north. These rocks were initially deposited in a broad shallow basin unconformably covering the autochthonous rocks (Bohannon, 1984). The Rainbow Gardens Member of the Horse Spring Formation represents the lower Tertiary section. The Rainbow Gardens includes clastic rocks

ranging in grain size from conglomerate to claystone, several types of carbonates, evaporites, and cherts. Later faulting disrupted this broad basin, and sedimentation of the upper Horse Spring Formation (the Thumb Member and above) occurred within smaller, fault-controlled basins (Bohannon, 1984). The upper Horse Spring includes clastic, carbonate, and tuffaceous rocks. The nearly unconsolidated Tertiary Muddy Creek Formation and Quaternary fanglomerates filled most of the fault-controlled basins, reaching thicknesses of at least 215 m in the Muddy and Virgin river valleys, and 425 m in Detrital Valley (Bohannon, 1984). The Muddy Creek Formation consists of siltstone, sandstone, gypsum, gypsiferous siltstone, and conglomerate. Tertiary and later sediments are thin or absent in the Black Canyon area, having been scoured away by the Colorado River (Anderson and Laney, 1975).

Unconsolidated Pleistocene or Recent alluvial deposits are composed of alluvial fan, fluvial, fanglomerate, lakebed, and aeolian deposits (Longwell *et al.*, 1965). Locally, coarse-grained Quaternary deposits are cemented with calcium carbonate. Older, moderately-well-cemented, fluvial deposits are exposed in the walls of Mormon Mesa, between the Virgin and Muddy Rivers.

One of the earlier periods of deformation that strongly affected the study area was the Sevier orogeny during late Cretaceous to early Tertiary. This event of eastward-directed thrust faulting disrupted the stratigraphic section, placing Paleozoic carbonates over Jurassic sandstones. One of the easternmost thrust systems is the Muddy Mountain thrust system which formed the Muddy Mountains located in the northern portions of the study area (Longwell, 1922).

During late Tertiary, major strike-slip and normal faulting associated with Basin and Range extension disrupted the Lake Mead area. Strike-slip faulting dominates the study area north of the lake and these late Miocene faults are known collectively as the Lake Mead fault system (Anderson, 1971). Comprised of numerous discontinuous left-lateral strike-slip faults, the Lake Mead fault system has an estimated total displacement of 60 km distributed along its entire length and fault segments (Bohannon, 1984). Two of these fault segments, the Bitter Spring Valley and the Rogers Spring faults, bound the Overton Arm pull-apart basin (Campagna and Aydin, 1994). Several large springs in the study area are located along the Rogers Spring fault near its southwestern terminus. There, the Rogers Spring fault separates the younger Tertiary through Quaternary sediments of the Overton Arm basin on the east from the allochthonous Paleozoic section of the Muddy Mountains on the west. In this area, the fault strikes N50°E, is vertical to 75°SE dipping, and has a gouge zone up to 5 m thick (Campagna and Aydin, 1994). Northeast of the Muddy Mountains, the Rogers Spring fault lies entirely within the Muddy Creek Formation, strikes N60°E, and is nearly vertical. The thickness of the zone of low-permeability fault gouge and the transition from transmissive carbonate rocks to low-permeability basin-fill sediments creates a barrier to further eastward flow of groundwater.

The extreme western portions of the study area include Frenchman Mountain, which is bounded by northwest trending right-lateral strike-slip faults of the Las Vegas Valley shear zone. Longwell (1960) first identified the Las Vegas Valley shear zone as a northwest-trending right-lateral strike-slip fault beneath the alluvial fill of Las Vegas Valley. One of the faults passes

north of Frenchman Mountain and terminates at or near the southwestern extension of the Lake Mead fault system (Cakir, 1990; Duebendorfer and Wallin, 1991). Other faults within the system continue southeast past Frenchman Mountain (Campagna and Aydin, 1994), presumably terminating at the River Mountains and McCullough Range.

Normal faults, characteristic of Basin and Range extensional deformation, are most common south of the lake. In the Black Canyon area, normal faults are associated with magmatism, strike North-South, and dip at high angles to the west and east (Anderson *et al.*, 1994). These high-angle faults may become listric at depth (Anderson, 1971), providing horizontal pathways for groundwater flow in the volcanic terrain (McKay and Zimmerman, 1983). In addition, numerous small faults in this area strike N50°W and are oblique right-lateral strike-slip faults (Anderson, 1971).

In summary, the most important stratigraphic units that shape the hydrogeologic setting are the thick Paleozoic carbonates in the northwest, the thick Tertiary sediments that fill structural basins in the north, and the Precambrian and Tertiary igneous and metamorphic rocks in the south. The most important structural features are the Lake Mead strike-slip fault system in the north, and the normal faulting in the south.

GROUNDWATER FLOW SYSTEMS

Regional Flow Patterns

Groundwater flow systems in the Basin and Range province range in size from small local systems to regional systems that extend over hundreds of kilometers. Local systems usually occupy a single topographic or hydrographic basin and have short flow paths relative to regional systems. Regional systems incorporate multiple topographic basins and therefore interbasin flow is important. While local systems may receive the majority of their recharge in the local topographic basin, regional systems typically receive recharge from multiple basins, and local recharge in any particular basin may be minimal.

Southeastern Nevada comprises the ultimate groundwater discharge location for much of the eastern portion of the regional carbonate aquifer (Dettinger *et al.*, 1995). Major groundwater flow systems comprised of thick carbonate rocks enter the area from the north and meet hydrogeologic barriers to flow, formed by thick, low-permeability Tertiary basin-fill deposits and a Precambrain terrain intruded by Cenozoic igneous rocks. Near these barriers, groundwater is discharged directly at regional springs, or by upward flow into basin-fill aquifers and subsequently discharged by evapotranspiration, spring flow, and streams. Groundwater flow in northwestern Arizona is less well-defined, but generally occurs as northward flow in the basin-fill deposits of Detrital Valley, with ultimate discharge to Lake Mead, and westward flow in basin-fill deposits, and perhaps igneous rocks, toward the Colorado River (Bedinger *et al.*, 1984). The generalized directions of groundwater flow in southeastern Nevada and northwestern Arizona are shown in Figure 5.

Most groundwater in the Basin and Range geologic province flows through carbonate-rock aquifers interconnected with unconsolidated basin-fill aquifers. In southern Nevada, basin-fill aquifers tend to be isolated by topographic divides and contribute to multi-basin groundwater flow

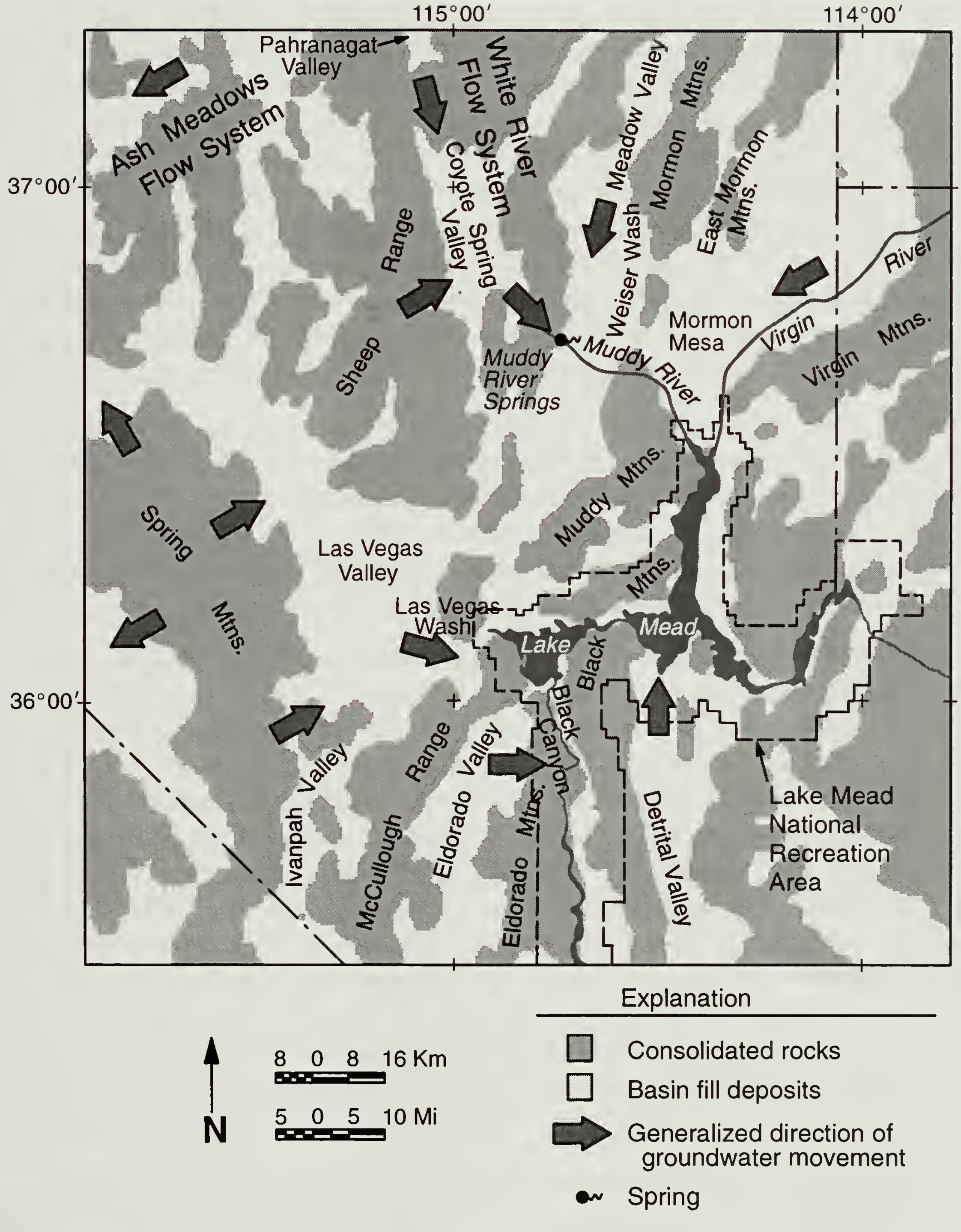

Figure 5. Regional groundwater flow patterns in southeastern Nevada and extreme northwestern Arizona. Modified from Harrill *et al.* (1988) and Bedinger *et al.* (1984).

systems only when they are in close hydraulic connection with underlying carbonate rocks. The most permeable basin-fill sediments were deposited as alluvial-fan, lake-bed, or fluvial deposits in basins formed by late Tertiary and Quaternary normal faulting. The earlier Tertiary basin-fill sediments of the Horse Spring and Muddy Creek Formations are generally less permeable due to finer grain size.

High transmissivities in carbonate rocks result from their great thickness, numerous faults and fractures caused by extensional deformation of the brittle carbonate rock, and to a lesser degree, solution enlargement of fractures and joints (Dettinger *et al.*, 1995). The high transmissivity of these rocks has been demonstrated during pumping tests in several wells, including the MX wells in Coyote Spring Valley (Bunch and Harrill, 1984) and the Arrow Canyon well in the Moapa Valley (Buqo, 1993), and by the high discharge rates from regional springs in the carbonate-rock province (Eakin, 1964). The carbonate rocks do not form a continuous unit, but rather are composed of many discrete structural blocks bounded by faults (Plume and Carlson, 1988). This pattern is manifested at land surface as distinct, often closed, topographic basins surrounded by mountain ranges. The transmissive carbonate rocks often provide a mechanism for deep groundwater flow between basins where topographic divides prevent shallow flow between adjacent basin-fill aquifers (Eakin, 1966).

Orographic effects cause most recharge within the carbonate rock province to be derived from precipitation in the higher elevations of east-central Nevada (Eakin, 1966). Groundwater recharge is minimal in low-elevation basins because potential recharge from precipitation is quickly lost to evapotranspiration (Maxey *et al.*, 1966). The carbonate aquifers of southern Nevada are recharged primarily from precipitation at high altitudes in the nearby Sheep and Spring Mountains (Winograd and Riggs, 1984), and from flow that enters the region from carbonate aquifers to the north.

Two major flow systems have been delineated within the southern part of the carbonate terrain. One discharges approximately 130 km west of the study area at Ash Meadows and Death Valley (Winograd and Thordardson, 1975) and the other, the White River flow system, discharges at the Muddy River Springs in the Moapa Valley (Eakin, 1966). The latter flow system is pertinent to any study of groundwater resources in southeastern Nevada because it supplies the vast majority of groundwater flow into the region. It comprises thirteen interconnected groundwater basins that extend over 370 km north to Long Valley (Eakin, 1968). The Muddy River springs are believed to be the primary regional discharge point of the White River System (Eakin, 1968), although groundwater from other basins, namely Meadow Valley, may contribute some discharge to the springs (Schroth, 1987; Kirk and Campana, 1988; Thomas *et al.*, 1997). In addition, Thomas *et al.* (1997) suggest that most groundwater recharge in the Sheep Range, which is located directly west, may be discharged at the Muddy River Springs. The Muddy River spring area represents the single greatest groundwater discharge point in southern Nevada, with estimated annual discharge of approximately 36,000 acre-ft/year (AFY) (Eakin, 1964; Prudic *et al.*, 1993; Thomas *et al.*, 1997).

Dettinger *et al.* (1995) summarize the evidence for the discharge at the Muddy River Springs and the related upward flow into overlying basin-fill aquifers in the area as being the terminus of the White River flow system. First, geologic constraints to the east and southeast of the Muddy River Springs suggest further flow in those directions and toward Lake Mead is unlikely. These constraints include the thinning of carbonate rocks and exposure of Precambrian crystalline basement rocks on the western edge of the Mormon Mountains; thick (over 1200 m), low-permeability basin-fill sediments just east of the springs below California Wash; and, except for isolated areas, few carbonate rocks extending below Lake Mead (Longwell, 1936). Second, Longwell's mapping of the floor of present-day Lake Mead revealed no evidence of spring discharge. Finally, spring

temperatures and stable isotopic data (to be discussed in more detail in a later section of this report) suggest that large down-gradient springs (Rogers and Blue Point springs near the Overton Arm of Lake Mead) are not directly related to discharge at the Muddy River Springs.

There is, however, evidence of groundwater discharge to the Muddy River about 20 km downstream of the Muddy River springs. Here, the Muddy River passes through "The Narrows" formed by the North Muddy Mountains and the Mormon Mountains. Rush (1968) reports gains in Muddy River discharge of 170 L/s in this reach and suggests that the most probable source for the flow is consolidated rocks underlying the thin alluvium. Although not discussed by Rush (1968), this discharge might represent the last point of discharge for flow from the White River flow system, or might represent flow from the Weiser Wash and Mormon Mountain regions directly north.

Another source of groundwater flow into southeastern Nevada is the Virgin River Valley to the northeast of the Overton Arm, although there is disagreement as to the amounts and locations of discharge. Glancy and Van Denburgh (1969) estimate groundwater discharge to Lake Mead through the valley fill and underlying consolidated rocks to be as much as 40,000 AFY. Most of this discharge was thought to be seepage from the Virgin River, which is a losing stream through much of the lower Virgin River Valley. However, Prudic *et al.* (1993) include no subsurface discharge from the Virgin River Valley to Lake Mead in their numerical model of regional groundwater flow. Instead, all groundwater in the near-surface aquifer is simulated as discharge by evapotranspiration (8000 AFY) or baseflow to the Virgin River (5000 AFY), while all discharge in the lower layer of the model (presumably consolidated rocks) is simulated as discharge at Rogers and Blue Point Springs (1200 AFY). The remainder of the discharge is considered surface flow in the Virgin River, and is not included in the model.

In the Las Vegas Valley, numerical modeling (Harrill, 1976; Morgan and Dettinger, 1994) and stable isotopic data (Thomas *et al.*, 1997) indicate that the majority of groundwater originates in the Spring Mountains to the west, with only minor amounts of recharge received from the Sheep Range. Thomas *et al.* (1997) suggest that structural constraints to the west, south, and southeast of the Sheep Range prevent groundwater flow in those directions, thus forcing flow toward Coyote Spring Valley to the northeast. Based on hydraulic head data, Thomas *et al.* (1997) suggest that a small amount of groundwater flow may also originate from Ivanpah Valley to the southwest, although, based on stable isotopic data, the southern portion of the Spring Mountains is the most important source of recharge to the southwestern portion of the Las Vegas Valley.

Hydraulic head relationships indicate that discharge from the Las Vegas Valley is to the east toward Lake Mead, although the amounts are likely to be small (Rush, 1968). Significant subsurface flow beneath Las Vegas Wash is unlikely because the basin fill below the channel is comprised of deposits of the low-permeability Muddy Creek Formation (Rush, 1968). Elsewhere, subsurface flow must pass through low permeability consolidated rocks and is therefore considered minimal. Calibration of numerical models (Harrill, 1976; Morgan and Dettinger, 1994) suggests less than 2000 AFY is discharged from the Las Vegas Valley toward Lake Mead in the area of Frenchman

Mountain. There exists little evidence for significant groundwater flow in the Tertiary volcanic rocks near Lake Mead (Laney and Bales, 1996).

The termini of groundwater flow systems in southern Nevada are located in areas where geologic constraints prevent further subsurface flow, causing discharge at the surface via springs and evapotranspiration; or where land surface elevations are sufficiently low to intersect groundwater flow paths. As previously described, the Muddy River Springs area is representative of the first mechanism, forming the terminus of the White River flow system and discharging approximately 36,000 AFY. The locations of Rogers and Blue Point springs, which have a combined discharge of approximately 1200 AFY (Laney and Bales, 1996), and other nearby springs, are also related to geologic constraints; that is, the transition from transmissive carbonate rocks to low-permeability basin-fill formed by the Rogers Spring Fault. Until recently however, the origin of groundwater discharged at these springs has been uncertain. Similarities between the geologic setting west of the Overton Arm and in the Moapa Valley lead early workers to group them with the Muddy River springs, making Rogers and Blue Point springs the terminal end of the White River flow system. Additional information about the physical, chemical, and isotopic nature of groundwater flow systems in southern Nevada has lead to new interpretations, including probable flow from the Virgin Valley to the north (Prudic *et al.*, 1993) and from recharge areas in the Sheep Range to the west and/or Mormon Mountains to the northwest (Dettinger *et al.*, 1995; Thomas *et al.*, 1997).

The Black Canyon of the Colorado River is suggested by Rush and Huxel (1966) and Mifflin (1968) as another discharge area within southern Nevada, primarily for the McCullough Range and Eldorado Valley. Evidence includes the presence of several springs and seeps at the base of Black Canyon near the present location of Hoover Dam that were noted during investigations for, and construction of, the dam (U.S. Bureau of Reclamation, 1950). The adjacent Black Mountains and Eldorado Mountains are suggested by McKay and Zimmerman (1983) as possible sources for several springs in Black Canyon, based on stable isotopic data that indicate low-elevation recharge. However, stable isotopic data for local precipitation and groundwater recharge were not available at the time of their study, and McKay and Zimmerman conclude that insufficient evidence existed for significant groundwater recharge at the low elevations in these areas. In addition, McKay and Zimmerman (1983) suggest that the permeability of faults and fractures in the volcanic rocks of Black Canyon is sufficient to provide important pathways for groundwater flow. Finally, McKay and Zimmerman (1983) provide strong evidence for the influence of recirculated Lake Mead water on several springs in Black Canyon.

Chemical Composition of Groundwaters

The limestone and dolomite that form carbonate aquifers are dominated by the soluble minerals calcite and dolomite, resulting in a calcium and magnesium-bicarbonate water composition that is fairly homogeneous throughout the carbonate-rock province of eastern and southern Nevada (Hess and Mifflin, 1978). Other minerals, such as gypsum and halite, are present in carbonate rocks in minute amounts but are more soluble than the carbonate minerals. Maxey and Mifflin (1966) show that solution of these minerals causes characteristic increases in the concentrations of the ions

sodium, potassium, chloride, and sulfate as groundwater moves along regional flow paths. Overall, the water quality in carbonate rocks in southern Nevada is generally good, with TDS concentrations less than 600 mg/L (Lyles *et al.*, 1987).

Hershey and Mizell (1995) demonstrate the evolution of groundwater chemistry in the carbonate flow system of southern Nevada using a trilinear plot of major dissolved ions in regional carbonate springs (Figure 6). The groundwater flow paths implied on this plot are based on regional flow patterns proposed by Harrill *et al.* (1988). Groundwater intermediate in the flow system is represented by springs in Pahranagat Valley and White River Valley (the next valley north and upgradient of Pahranagat Valley) which show the calcium, sodium-bicarbonate and sulfate composition typical of carbonate waters. One evolutionary trend follows the flow path toward the regional discharge point at the Muddy River Springs. Groundwater flow along this path is

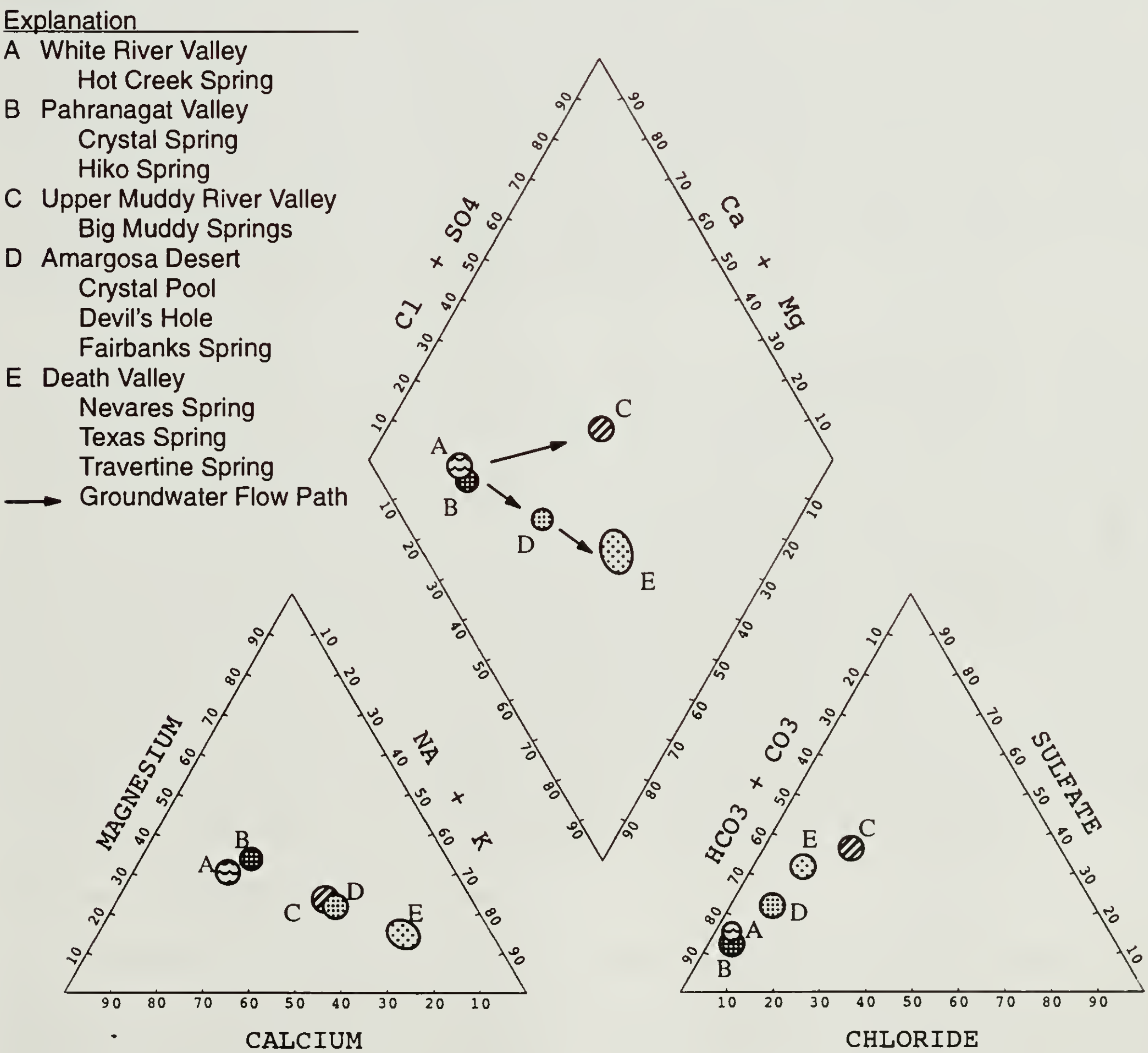

Figure 6. Trilinear diagram showing major dissolved ions of regional springs in the carbonate-rock province of eastern Nevada, showing evolution of groundwater chemistry along two flow paths. Modified from Hershey and Mizell (1995).

accompanied by increases in the concentrations of sodium, potassium, sulfate, and chloride ions attributed to solution of evaporite minerals in the Horse Spring and Muddy Creek Formations near the discharge point. Calcium and magnesium also increase, but to a lesser degree. The other evolutionary trend follows a flow path through Ash Meadows to the regional discharge point in Death Valley. Increases in the concentrations of all major ions except calcium and magnesium along this flow path to Ash Meadows are attributed to solution of Tertiary silicic volcanic rocks (Winograd and Thordardson, 1975). From Ash Meadows to Death Valley, concentrations of all major ions except calcium and magnesium increase as a result of solution of Tertiary and Quaternary lacustrine and alluvial deposits. Declines in the concentrations of calcium and magnesium are attributed to cation-exchange with clays and precipitation of travertine deposits at the springs.

Groundwater in volcanic rocks northwest of Las Vegas is generally of sodium and potassium-bicarbonate composition, reflecting dissolution of feldspar and mafic minerals along relatively long flow paths (Winograd and Thordardson, 1975; Lyles *et al.*, 1987). Locally, waters collected from springs south of Las Vegas in the McCullough Range and a well and springs in the Eldorado Mountains have a mixed cation-sulfate or a mixed cation-bicarbonate composition (Lyles *et al.*, 1987; SNWA, unpublished data) (Figure 7), similar to springs that represent early-stage recharge chemistry in volcanic rocks of central Nevada (Raker and Jacobson, 1987). TDS concentrations of the McCullough Range samples range from 414 mg/L to 664 mg/L while the Eldorado Mountains samples range from 957 mg/L to 1390 mg/L.

Groundwater in basin-fill deposits is categorized as calcium and magnesium-bicarbonate, mixed cation-sulfate, and sodium and potassium-bicarbonate composition (Figure 7) (data from Lyles *et al.*, 1987). Composition varies considerably across the region, depending on lithology, residence time, and origin. Groundwater quality is poorest in the eastern portion of the region, and is characterized by TDS concentrations that range from about 1000 mg/L to well over 2000 mg/L, and mixed cation-sulfate composition (Lyles *et al.*, 1987). The sulfate is derived from solution of evaporite minerals, including gypsum and thenardite (Lyles *et al.*, 1987), in sedimentary rocks of Tertiary age (Muddy Creek and Horse Springs Formations), Triassic age (Moenave, Kayenta, and Moenkopi Formations), and Permian age (Permian Red Beds and Kaibab-Toroweap Formations) (Bohannon, 1984). These rocks are abundant at the surface and in the near surface from Frenchman Mountain northeast to the Overton Arm, and commonly overlie, or are structurally adjacent to, Paleozoic carbonate rocks. Thus, groundwater in this area is likely to pass through evaporite deposits at some point along flow paths, greatly increasing TDS and sulfate concentrations.

Isotopic Composition of Groundwaters

Groundwater in southern Nevada is derived from two principal sources: recharge from local precipitation, and groundwater flowing into the area from regional and subregional aquifer systems described above. Groundwater recharge can be further divided into recharge at altitudes less than 1500 m, which includes most of the region; and recharge at altitudes above 1500 m, which in southern Nevada is limited primarily to the Spring Mountains and Sheep Range. Although smaller in area and lower in altitude than these ranges, the Mormon Mountains also receive precipitation

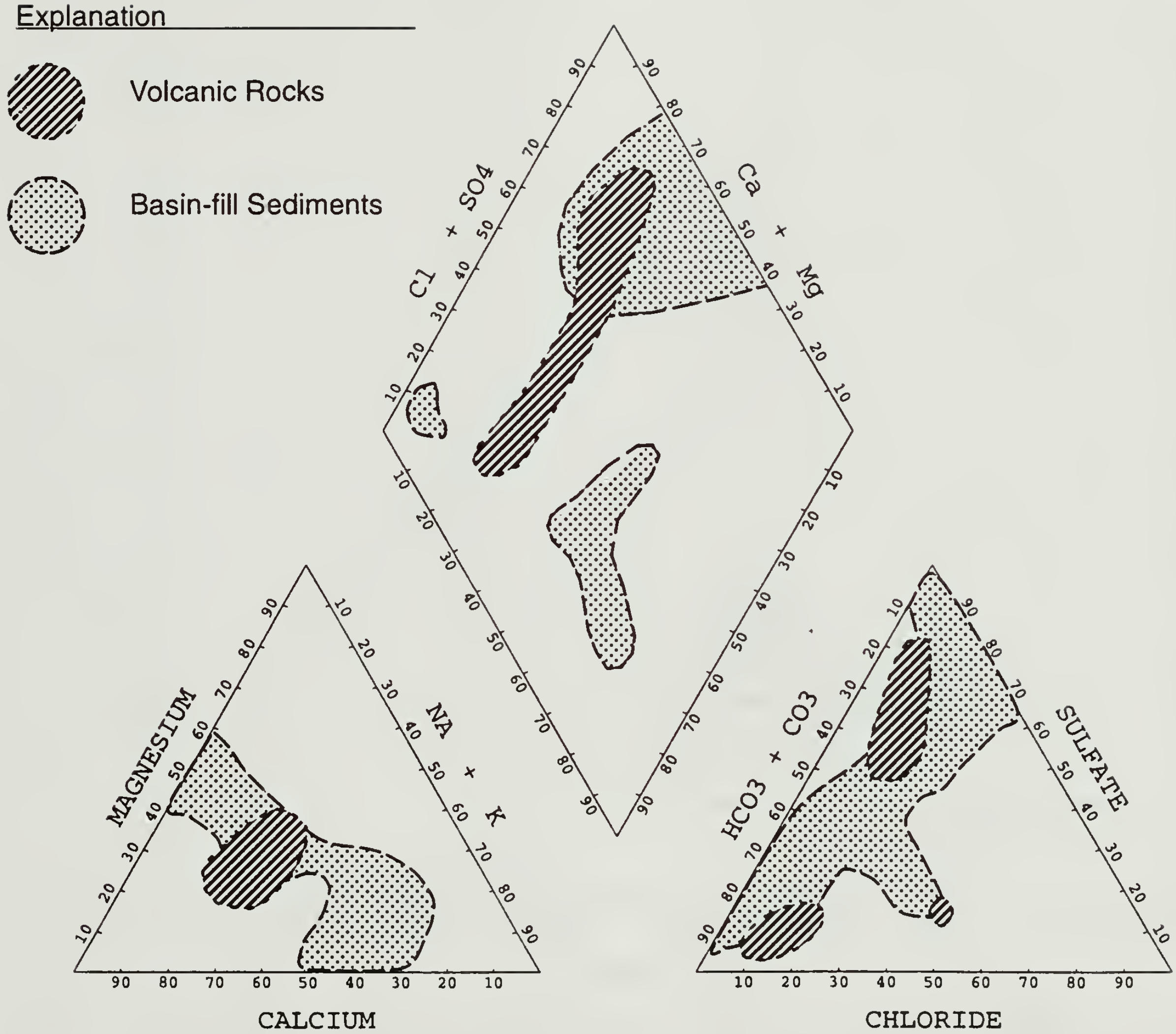

Figure 7. Trilinear diagram showing major dissolved ions in groundwaters collected from volcanic rocks and basin-fill sediments in southern Nevada. Modified from Lyles *et al.* (1987), with additional data from SNWA (unpublished data).

at altitudes above 1500 m and are located much closer to Lake Mead. Although precipitation is the ultimate source of groundwater recharge, evaporation and associated isotope fractionation during recharge under arid conditions causes recharge waters to have a different isotopic composition than the original precipitation. Therefore, selected spring data are used in the present study to represent the stable isotopic composition of groundwater recharge. In addition, local precipitation data were not available at the time of the present study and the timeframe of the study did not allow for long-term precipitation collection.

The stable isotopic values of springs in selected groundwater recharge areas are shown in Figure 8. Also shown is the global Meteoric Water Line (MWL) that represents the linear relationship between $\delta^{18}O$ and δD described by Craig (1961) using data from over 400 rivers, lakes, and precipitation. The local MWL shown represents precipitation (falling as rain) at 32 sites in

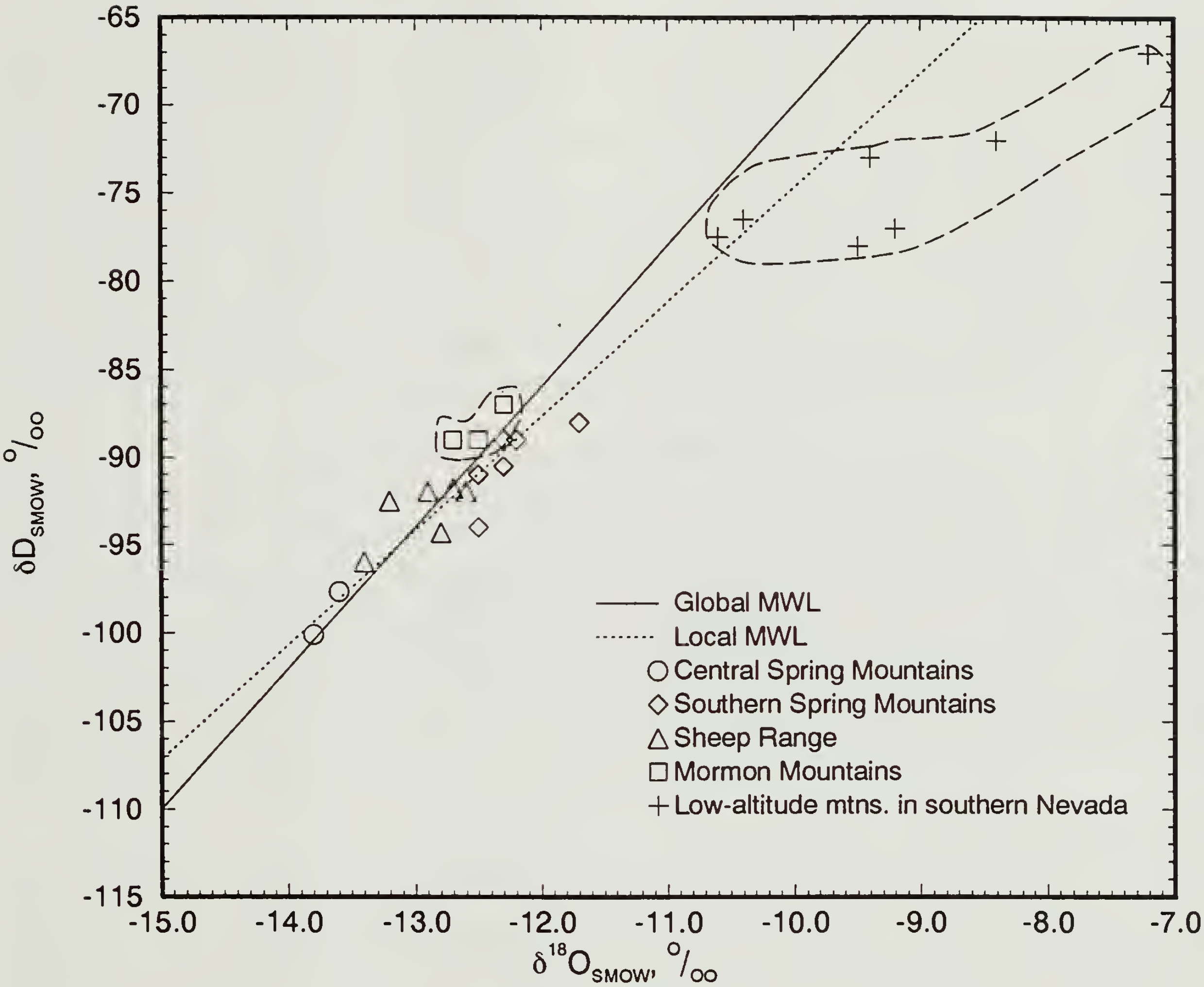

Figure 8. Stable isotopic composition of springs in groundwater recharge areas in southern Nevada, the global meteoric water line (after Craig, 1961), and a local meteoric water line (see text for description). Groundwater recharged at high altitudes in the Spring Mountains, Sheep Range, and Mormon Mountains is isotopically lighter than groundwater recharged in low-altitude mountain ranges. Data are compiled in Table C-1.

southeastern California between April 1986 and October 1987 (Friedman *et al.*, 1992). The equation for the least squares line for this data set is $\delta D=6.5\delta^{18}O-9.7$. Springs located at altitudes above 1100 m in the Spring Mountains, Sheep Range, and Mormon Mountains plot as the isotopically lightest points on Figure 8 (data from Thomas *et al.*, 1997). As atmospheric moisture rises up the mountain fronts, the heavier isotopes of hydrogen and oxygen are selectively removed with precipitation and the residual moisture becomes isotopically lighter. Thus, groundwater recharged at high altitudes in these mountains is isotopically lighter than groundwater recharged at lower altitudes. Thomas *et al.* (1997) also note that the higher altitudes of the central Spring Mountains result in more depleted stable isotopic compositions compared to the southern Spring Mountains. Springs in both portions of the Spring Mountains also contain tritium concentrations of up to 257 pCi/L (analyzed in 1976), indicating a major component of post-1952 recharge. The $\delta^{13}C$ concentrations range from -7.9 to -11.2 per mil, reflecting the enrichment of $\delta^{13}C$ by dissolution of carbonate rocks.

The isotopically heavier, low altitude points shown on Figure 8 represent springs that are derived from recharge that occurs at altitudes less than 1500 m in the McCullough Range and Eldorado Mountains adjacent to Black Canyon, the Highland Range and New York Mountains south of Eldorado Valley, and the East Mormon Mountains northwest of Lake Mead (Thomas *et al.*, 1997; SNWA, unpublished data). These springs are located in ranges, or in portions of ranges, that receive most of their recharge at altitudes lower than about 1500 m. Because they are located at altitudes above the adjacent valleys, and therefore are unrelated to regional groundwater flow systems, groundwater discharged from these springs represents local, low-elevation recharge rather than regional groundwater flow. The existence of these springs indicates that local recharge can be more significant than basin-wide predictions developed using the Maxey-Eakin method.

The greater spread of the low altitude data points on Figure 8 likely results from local differences in conditions and seasons of recharge in each individual spring catchment area. The isotopic composition of these springs is reasonably consistent with precipitation data collected at Searchlight, Nevada between the years of 1982 to 1989 (average annual δD of -73 per mil) (Friedman *et al.*, 1992) and at the Nevada Test Site (average annual δD of -80 per mil) (Ingraham *et al.*, 1991). Therefore, the stable isotopic composition of local, low-elevation recharge in the area of study is assumed to be that of these low-elevation mountain springs. It should be noted that these springs plot close to the estimated composition of present-day groundwater recharge near Searchlight, Nevada (δD of -80 per mil) (Smith *et al.*, 1992).

Tritium data for the low-elevation springs are sparse. However, tritium values have been measured at two springs in the Eldorado Mountains. These concentrations (19 and 24 pCi/L – analyzed in 1995; SNWA, unpublished data) indicate that post-1952 recharge contributes to flow at these springs. The only ^{14}C data available for low-elevation springs is for a single spring in the McCullough Range. This spring contains 68.1 percent modern carbon (PMC), for an uncorrected age of 3,175 years, which further distinguishes it from older, regional groundwater flow.

The stable isotopic composition of groundwater in regional and subregional flow systems is shown in Figure 9. Data from the White River flow system of the regional carbonate aquifer (Thomas *et al.*, 1991; DRI, unpublished data) show a trend toward heavier composition along the flow path from Pahranagat Valley (white triangles), through Coyote Spring Valley and other nearby valleys (light shaded triangles), to the Muddy River Springs (dark triangles). Groundwater is isotopically lightest at the recharge areas in east-central Nevada, where recharge occurs at higher elevations and under different climatic conditions, and becomes isotopically heavier as local, lower-elevation precipitation recharges the system. Between Pahranagat Valley and the Muddy River Springs, the addition of isotopically heavier groundwater originating from the Meadow Valley flow system to the northeast, and recharge in the Sheep Range to the west is thought to cause the composition observed at the Muddy River Springs (Kirk and Campana, 1988; Thomas *et al.*, 1997).

Tritium is below detection levels in the southern part of the regional carbonate aquifer (Hershey and Mizell, 1995), reflecting long travel times from recharge areas and/or the dilution of local recharge with regional flow. In addition, the carbonate system shows trends of decreasing PMC and

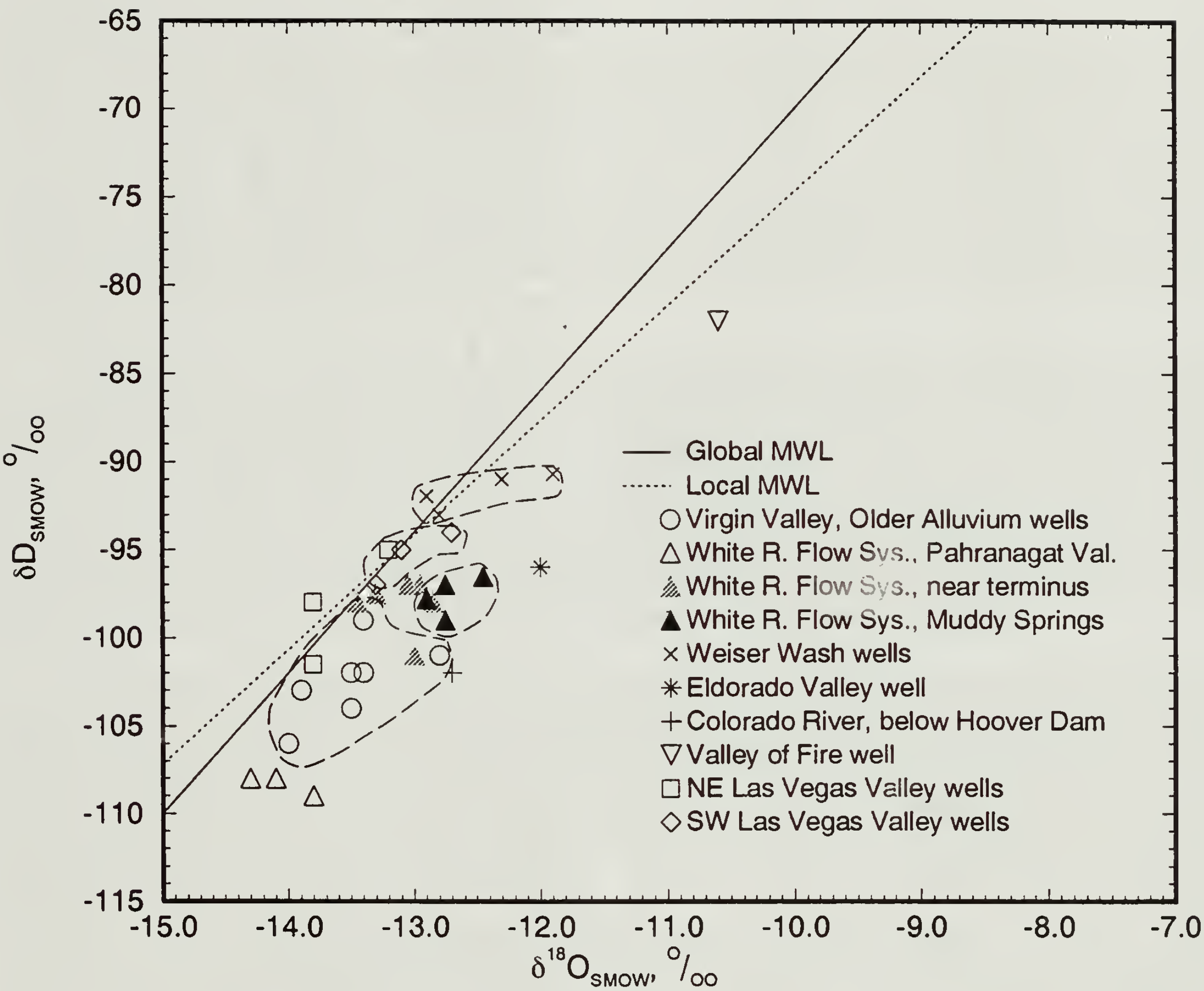

Figure 9. Stable isotopic composition of selected groundwaters of southern Nevada. Data are compiled in Table C-1.

increasing $\delta^{13}C$ values along regional flow paths, reflecting the increasing age of groundwater and dissolution of carbonate minerals with the addition of dead carbon and enrichment of $\delta^{13}C$ (Hershey and Mizell, 1995). Known regional springs in the carbonate aquifer system have PMC values of 2.8 to 11.2 and $\delta^{13}C$ values of -5.8 to -3.9 per mil (Hershey and Mizell, 1995).

In the Virgin Valley, groundwater obtained from wells in the Older Alluvium (which includes the Tertiary Muddy Creek Formation) is isotopically lighter than groundwater in the near surface aquifers and the Virgin River, suggesting a different origin (Metcalf, 1995). The composition is similar to that of groundwater at the southern end of the White River flow system, which may reflect a similar recharge source for Older Alluvium waters, such as carbonate aquifers to the north of the Virgin Valley (Glancy and Van Denburgh, 1969).

Isotopic similarities between groundwater in the basin-fill of Eldorado Valley and of southwest Las Vegas Valley, suggest a common origin. In addition to their similar δD values (Figure 9), groundwater in these two areas share $\delta^{13}C$ values between -6.8 and -7.8 which suggests flow in

carbonate rocks (or possible reactions with pedogenic carbonates or carbonate dust). Furthermore, PMC values of 7.75 and below are similar to the older, regional groundwaters noted above. Finally, the lack of detectable tritium in the Eldorado Valley sample suggests a pre-1952 age. Thomas *et al.* (1997) propose that groundwater in southwest Las Vegas Valley originates from low elevation recharge in the southern Spring Mountains. Although only a single data point is available in Eldorado Valley, the similarity to groundwater in southwest Las Vegas Valley is consistent with the idea of interbasin groundwater flow into and through Eldorado Valley, as proposed by McKay and Zimmerman (1983) and Harrill *et al.* (1988).

Groundwater of a more local, low-elevation origin is found in the Weiser Wash area, between the Mormon Mountains and the Muddy River. Here, water in the Muddy Creek Formation and underlying rocks is isotopically heavier (DRI, unpublished data) than groundwater in the Older Alluvium of Virgin Valley and groundwater discharged at the Muddy River Springs. This groundwater may represent a mixture of groundwater from the Meadow Valley Wash flow system (described by Thomas *et al.*, 1997) and isotopically heavier recharge (average δD of -88 per mil) in the Mormon Mountains. Tritium and carbon data are not available for this area.

Groundwater that appears to have a major component of locally-derived recharge occurs at Valley of Fire State Park, where a sample collected from the headquarters well has a heavier isotopic composition than most other groundwater in the region. Although the hydraulic head measured in this well conforms to the regional hydraulic head gradient between the Muddy River springs and Lake Mead, this area may represent a groundwater cell receiving local recharge through the Mesozoic sandstone terrain that covers the area. The $\delta^{13}C$ value of -8.5 per mil might represent reactions with pedogenic carbonates or carbonate dust, or might suggest a portion of the groundwater flows through carbonate rocks. The PMC value of 18.7, which is at least twice that of the upgradient Muddy River springs, indicates the presence another source of modern carbon.

Colorado River water (collected just below Hoover Dam) is isotopically lighter than most groundwater in the region, reflecting the isotopically-depleted composition of precipitation at higher elevations and cooler climates in the upper Colorado River drainage basin. The tritium concentration was 51 pCi/L in a water sample collected in 1997.

A selected set of uranium data for groundwaters in the region, including Rogers and Blue Point springs, is shown in Figure 10. This plot displays the $^{234}U/^{238}U$ activity ratio (AR) as a function of the inverse of total uranium concentration (μg/L). This plot reveals that there is an inverse relationship between uranium concentration (note that the x-axis is the reciprocal of concentration, so high concentrations plot to the left, and low concentrations plot to the right) and $^{234}U/^{238}U$ AR. This relationship has been widely observed, and has often been attributed to a trend line which shows evolution along a flowpath. According to this scenario, AR increases with the time that water has in contact with the aquifer matrix, and the concentration decreases as groundwater moves deeper along a flowpath, because it encounters reducing zones which causes uranium to precipitate from solution (Osmond and Cowart, 1992). Obviously, this scenario does not apply to waters of the region examined during this study, as no reducing zone is known to exist, even at great depths below ground

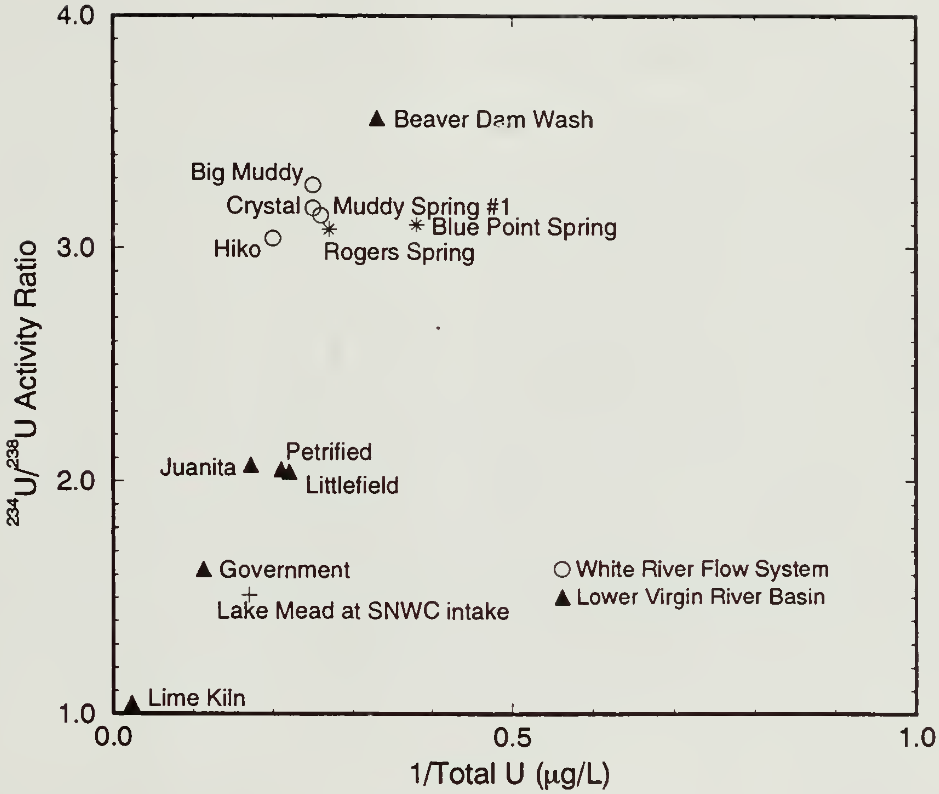

Figure 10. Uranium data for selected groundwaters of southern Nevada and southwestern Utah. Data compiled from Yelken (1996) and Farmer (1996).

surface. Kronfeld *et al.* (1994) present a scenario applicable to the deep oxygenated waters of the Basin and Range province, based on their study of an oxygenated carbonate aquifer in South Africa. The aquifer was found to exhibit a similar trend line to that seen in the present study, with the AR inversely related to concentration. Using ^3H and ^{14}C to date waters collected along the flowpath of the aquifer, the AR was shown to increase with the age of the water, indicating that a water moving along a flowpath would evolve from a low AR-high concentration signature to a high AR-low concentration signature with increasing residence time in the aquifer. Rainwater, which is typically very dilute, will begin to leach uranium from the soil and rock materials it encounters while recharging and flowing through an aquifer. As water flows through the aquifer, its AR increases, because ^{234}U, which is produced by alpha decay, is preferentially introduced due to a process called "alpha-recoil". Alpha-recoil enrichment is the result of the alpha decay process, which damages mineral crystal lattices in which decay occurs, making decayed mineral grains more susceptible to leaching than undamaged crystal sites; in other cases, the product of a decay can be injected directly into the liquid phase (Osmond and Cowart, 1994). Kronfeld *et al.* (1994) show that the uranium concentration in an oxygenated carbonate aquifer declines as water moves along the flowpath as a result of extensive ion exchange and/or sorption reactions with the aquifer matrix, with the ion exchange scenario appearing more likely. Based on these results, uranium data from this study may be interpreted such that waters with low ARs and high concentrations have had relatively short

periods of interaction with aquifer materials, and waters with high ARs and low concentrations have experienced relatively long periods in contact with aquifer materials.

The data shown in Figure 10 are derived from two primary areas - the lower Virgin River basin (Yelken, 1996) and the White River flow system (Farmer, 1996). Rogers and Blue Point Springs plot near members of the White River flow system, suggesting that waters from these two systems flow through rocks of similar type, and may have similar residence times. Waters in the Lower Virgin River Basin exhibit a wide rage of values, with outlier values suggesting both short and long residence times. The majority of these values are positioned so as to indicate intermediate travel times, suggesting that the springs in this region discharge waters having relatively short to intermediate residence times. These data support the classification of springs in the Virgin Mountains (Lime Kiln, Government, and Juanita Springs) as locally-derived, based on geographic considerations and stable isotope composition (Metcalf, 1995). Intermediate residence times for Petrified and Littlefield Springs (adjacent to the Virgin River, northeast of Mesquite, Nevada) support Metcalf's (1995) conclusion, based on stable isotope data, that these springs are not entirely locally-derived. The high AR of the sample from Beaver Dam Wash indicates a long residence time, suggesting that regional groundwater flow may form a significant component of baseflow to the wash.

RESULTS AND DISCUSSION

The physical, chemical, and isotopic data derived from previous studies, and collected for the present study, are compiled in Appendix A.

Spring Classification

For the purpose of the following discussion, the thirty-one springs in the Lake Mead National Recreation Area and the five nearby springs are divided into three sets based on the geographic nature of their source areas: local springs, subregional springs, and springs derived from Lake Mead water. Local springs discharge groundwater from small flow systems that receive most or all of their recharge locally and at low altitudes. Many of these local flow systems are contained entirely within the park boundaries. Subregional springs are dominated by groundwater that originates outside local topographic basins and flow systems, and may include groundwater recharged at higher altitudes. Most of the groundwater systems supplying the subregional springs extend beyond the park boundaries. In southern Nevada, a "regional" groundwater system generally denotes one that is part of the multi-basin carbonate aquifer system that extends over hundreds of kilometers. The term "subregional" is used here to avoid confusion. A third set of springs is derived from recirculated Lake Mead water.

Springs within each of the three sets share similar hydrogeologic settings and stable isotopic compositions, while discharge rates, temperatures, and tritium concentrations generally show considerable overlap. The distinct D and ^{18}O compositions of groundwater source areas and flow systems in southern Nevada makes the use of stable isotopes ideal for relating springs to their associated recharge sources. The following discussion will therefore focus on the hydrogeologic

settings and the isotopic compositions of springs as they relate to spring source areas. Discussion of the uranium data, which are available only for springs in the Lake Mead basin, follows in a separate section.

Local Springs

Lake Mead Basin

Six springs in the Lake Mead basin are considered local springs (Springs 1, 15, 16, 17, 18, and 19). Other than Spring 15, these springs have the lowest discharge rates in the Lake Mead basin, and their temperatures are strongly influenced by fluctuations in ambient air temperature (Table A-1 shows the pronounced differences between temperatures measured in October and February at these springs). These springs are not related to major structural features in the region, instead issuing from stratigraphic contacts, small faults or fractures, or simply at the intersection of the water table with land surface. With the exception of Spring 1, which issues from Quaternary terrace deposits at the base of Mormon Mesa, these springs discharge from alluvium or consolidated rocks in wash channels, and all support varying degrees of vegetation at their orifices. Evapotranspiration is a major controlling factor on the flow rate from these low-discharge springs, as evidenced by the variation in discharge observed at several springs between the seasons and time of day. Although a systematic study was not possible during the present investigation, flow rates at several low-discharge springs were highest during the winter months, and in the early morning hours during the summer months, when evapotranspiration rates of the vegetation surrounding the orifice are low. Discharge rates at these same springs was observed to be lower in the middle of the day in the summer months, when evapotranspiration rates are high.

Local springs in the Lake Mead basin exhibit a mixed cation-sulfate composition (Figure 11). Despite relatively short groundwater flow paths, these springs all have TDS values that exceed 1,200 mg/L. The high sulfate and TDS concentrations both originate from solution of the evaporite minerals so ubiquitous to the Permian, Triassic, and Tertiary rocks of the Lake Mead region.

The stable isotopic compositions of Springs 1, 15, 16, 17, 18, and 19 support their geographic and geologic designations as local springs. The stable isotopic compositions resemble local, low-elevation precipitation, especially if more depleted winter precipitation (Ingraham *et al.*, 1991) is considered (Figure 12). Springs 16 and 17 are located at altitudes above regional hydraulic heads, thus they may extend our definition of local recharge to more depleted δD values of -80 per mil. Additionally, these springs are significantly enriched in heavy isotopes compared to regional groundwater, indicating no relation to groundwater flow systems outside the study area.

The recharge areas for Springs 16 and 17 lie entirely within the park boundaries, in the Black Mountains area. The other local springs in the Lake Mead basin are recharged at least in part outside the park boundaries. Springs 1 and 15 lie on or near the eastern boundary, and their recharge areas extend outside the park. Recharge to Spring 15 originates within Bitter Spring Valley and White Basin, with possible contributions from the surrounding Muddy Mountains and other nearby, low-elevation areas. The δD composition of Spring 1 (-81 per mil) falls midway between the average

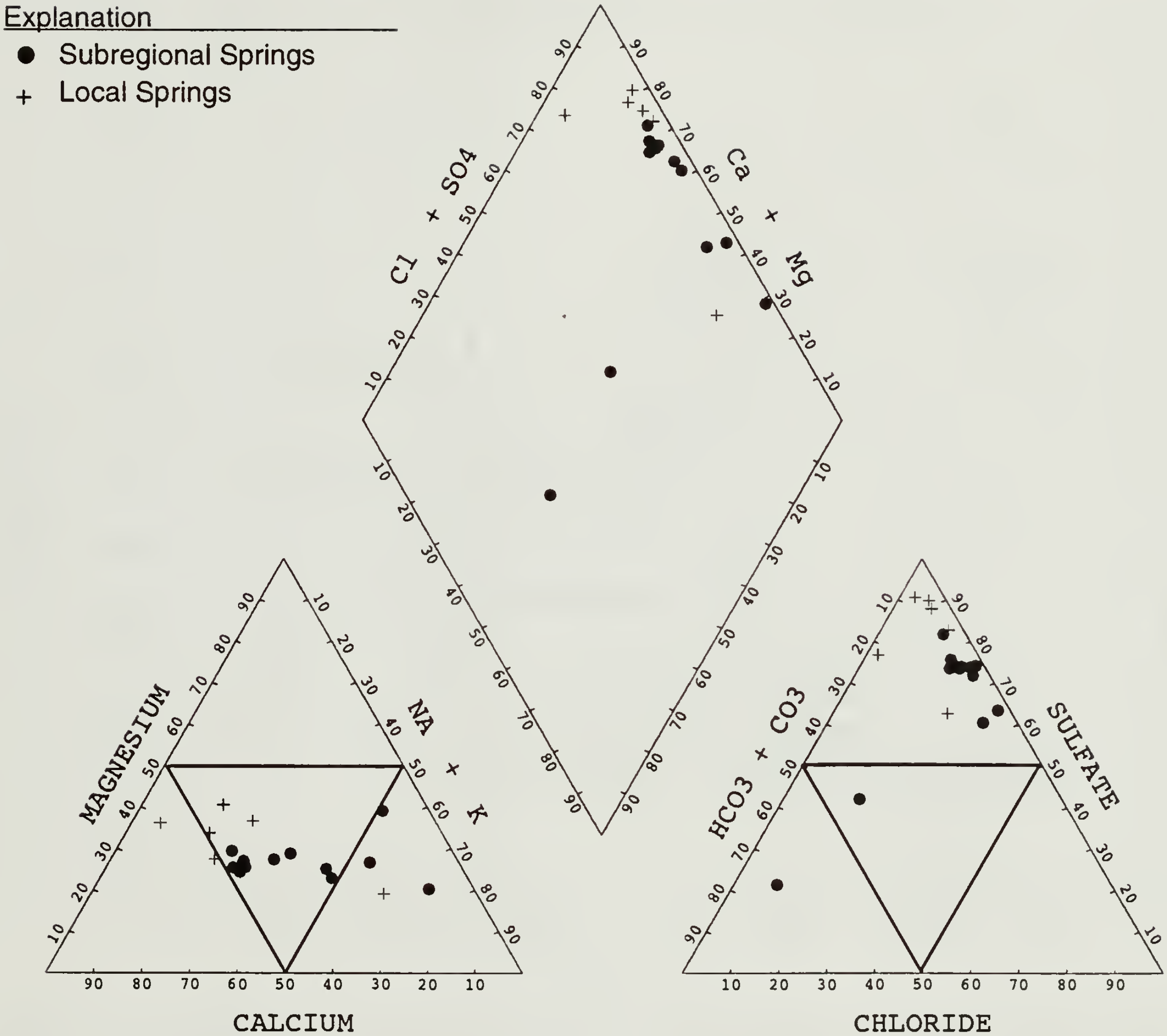

Figure 11. Trilinear diagram showing major dissolved ions of springs in the Lake Mead basin.

δD compositions of springs in the Mormon Mountains and springs in the East Mormon Mountains. Thus, it appears likely that flow from Spring 1 originates in the Mormon and East Mormon Mountains to the north, and travels through the alluvium that forms the upper portion of Mormon Mesa.

With the exception of Spring 1, local springs in the Lake Mead basin issue from alluvium or consolidated rocks in wash channels. However, the absence of atmospheric tritium indicates that groundwater travel times are long and that spring flow does not simply represent discharge of groundwater recharged during recent precipitation events.

Spring 18 appears to be controlled by the intersection of the water-bearing unit with land surface; as no structural control is evident. This spring plots in the region of low-elevation recharge which indicates that its flow originates locally. Although the elevation of the spring is lower than water levels in the carbonate aquifer to the north in Dry Lake Valley and to the west in the Las Vegas

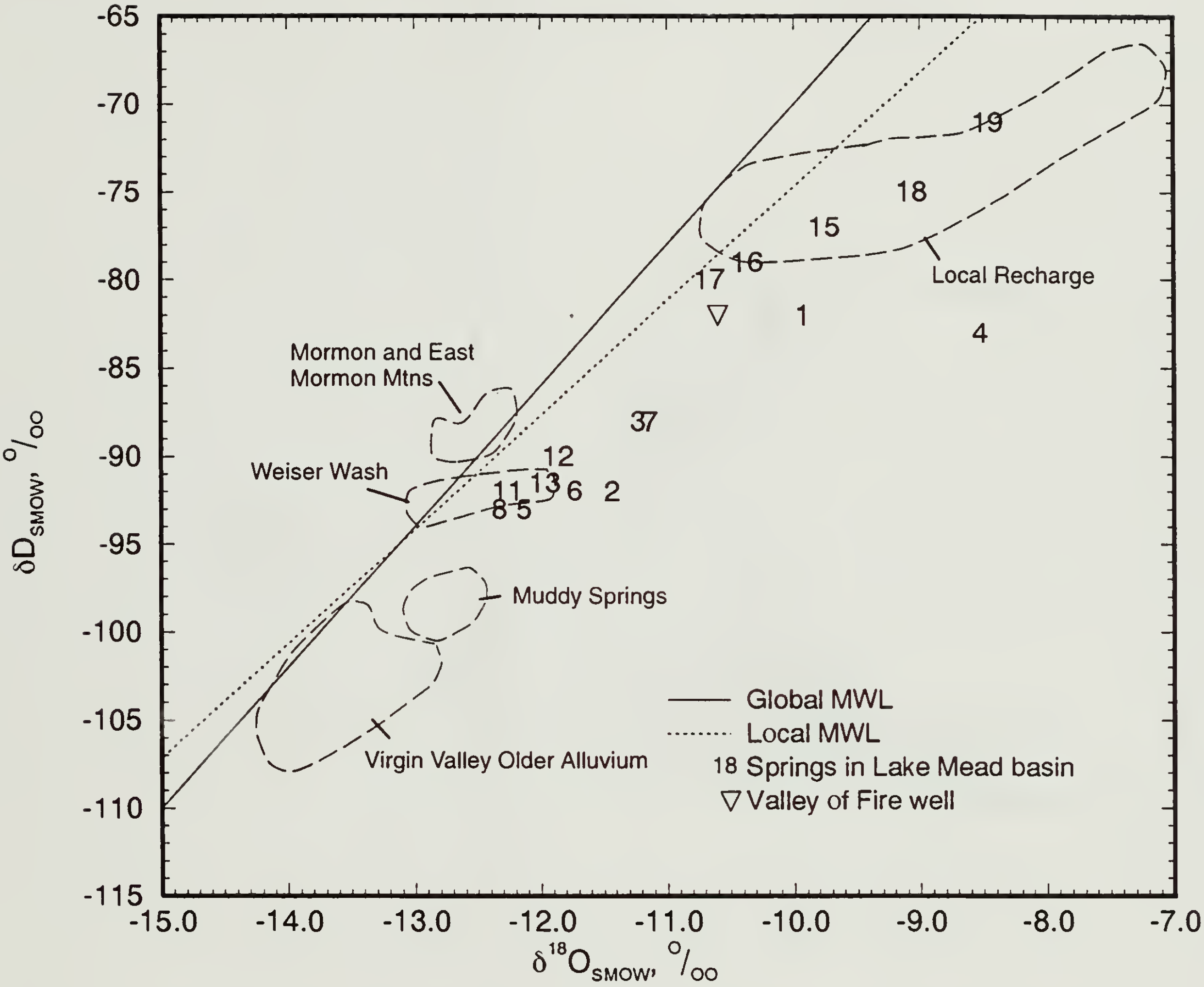

Figure 12. Stable isotopic composition of springs in the Lake Mead basin, as compared to other waters in the region.

Valley, the much more enriched δD composition of -75 per mil indicates that neither the carbonate aquifer nor Las Vegas Valley aquifers are the source.

<u>Black Canyon</u>

Three springs (Springs 27, 35, and 36) in the Black Canyon area are considered to be entirely of local origin. Discharge rates from these springs are less than 2 L/min, temperatures are less than 25 °C, and though these springs issue from alluvium in wash channels, their flow appears to originate from small faults or fractures in the underlying rock. Springs 27 and 36 are located at altitudes above regional hydraulic head in the Black Mountains and Eldorado Mountains, respectively, and their stable isotopic compositions fall within the region of low-elevation recharge on a plot of δD as a function of $\delta^{18}O$ (Figure 13). Spring 35 is located at a much lower altitude (960 m) and might be thought to be related to subregional flow; however, the stable isotopic composition clearly indicates local origin.

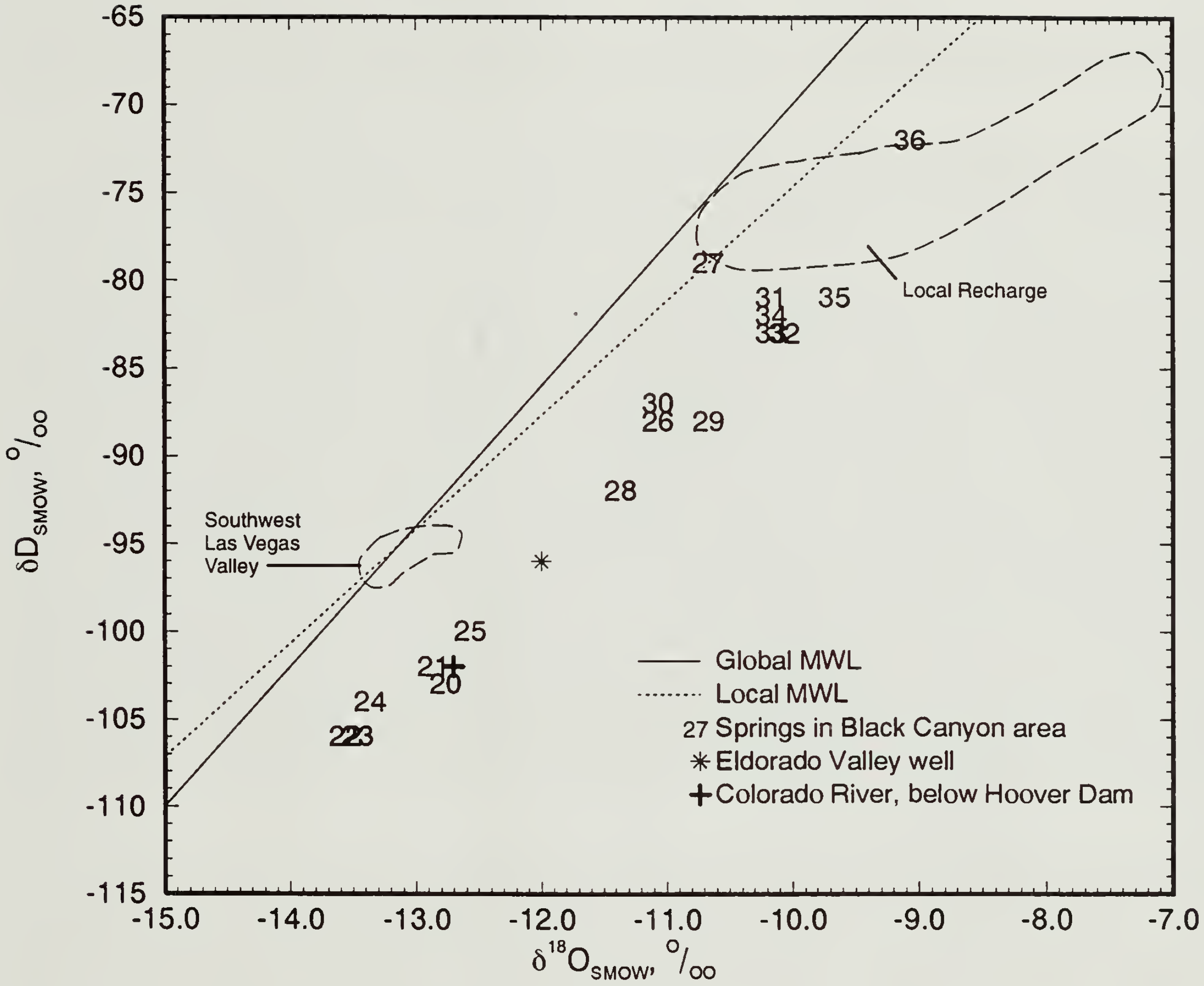

Figure 13. Stable isotopic composition of springs in the Black Canyon area, as compared to other waters in the region.

Unlike other locally-derived springs in the study area, Springs 27, 35 and 36 contain small quantities of detectable atmospheric ^3H (8.0, 8.2, and 18 pCi/L, respectively), indicating part of their discharge was recharged from precipitation after 1952. The fact that these springs contain atmospheric ^3H, while springs in the Lake Mead basin do not, may result from differences in the morphology of recharge catchment areas, and/or reflect infiltration of more recent precipitation in the alluvium upgradient of the springs. (Springs 27, 35, and 36 were sampled approximately one year after the others).

Four other springs in the Black Canyon area (Springs 31, 32, 33, and 34) are considered to be locally derived, but unlike Springs 27, 35, and 36, these springs are located in or near the bottom of Black Canyon. These springs range in distance from 6 to 11 kilometers south of Hoover Dam, and with the exception of Spring 31, issue directly from small faults in volcanic rock. Spring 31 issues from alluvium in the base of a wash channel, immediately upstream of where the channel becomes incised in volcanic bedrock. The discharge rates of these springs are higher than most of

the other locally-derived springs in the study area, ranging from less than 1 L/min to 10.2 L/min. Temperatures also tend to be higher, ranging from 19° to 32°C, reflecting the geothermal influence of intrusive rocks in the region (McKay and Zimmerman, 1983).

Springs 31, 32, 33, and 34 have virtually identical stable isotopic compositions (Figure 13), suggesting very similar conditions of groundwater recharge, despite the fact that two of the springs are located on the east side of the Colorado River and two are located on the west. The location of these springs at low altitudes near the groundwater discharge zone of the Colorado River suggests a potential relation to subregional flow, represented on the Nevada side by Eldorado Valley groundwater, and on the Arizona side by Detrital Valley groundwater. However, their stable isotopic compositions are very similar to local, low-elevation recharge, and are much more enriched in heavy isotopes than the Eldorado Valley well sample (δD composition of -96 per mil). These springs are slightly isotopically lighter than most of the other locally-derived springs, although the δD difference between them and locally-derived Spring 27 is only 3.3 per mil. Though this could result from different conditions of recharge, mixing of local precipitation with isotopically light subregional groundwater could also account for the isotopic composition and would be consistent with these springs' elevation, temperature, and flow rates. Due to their proximity, the Eldorado Mountains and Black Mountains represent the most likely sources of local, low-elevation recharge for these springs. Recharge from the McCullough Range, or other more distance ranges appears less likely due to the absence of any evidence of mixing with subregional groundwater (e.g., Eldorado Valley).

Though Spring 34 has a $\delta^{13}C$ composition similar to that of many other springs (-7.0 per mil, indicating a dissolved carbonate mineral contribution), Springs 31 and 33 are more unique, with their lighter carbon compositions (-13.2 and -24.9 per mil, respectively) indicating less contact between the groundwater and solid carbonate phases. For Springs 31 and 33, this suggests recharge through poorly developed soil and flow through strictly igneous terrain. A $\delta^{13}C$ value is not available for Spring 32. Considering the similar geologic settings of Springs 31, 33, and 34, the differences between their $\delta^{13}C$ values are not well understood at this time.

The absence of detectable atmospheric tritium in Springs 31, 32, 33, and 34 indicate that groundwater travel times are long and that these springs do not simply represent discharge of groundwater recharged during recent precipitation events. Limited ^{14}C data confirm this, but indicate widely varying apparent ages from 1660 to 15,500 years. Groundwater travel times from recharge areas to the springs of several thousands of years are consistent with their "local" designation and the arid environment. However, the age of 15,000 years obtained for Spring 33 seems inconsistent with other evidence of local origin, and indicates a more complex hydrochemical history than assumed here.

Subregional Springs

Lake Mead Basin

The majority of the springs studied in the Lake Mead basin are considered to be subregional springs. Most of these springs are located along North Shore Road, and as a group are termed the

North Shore Complex. These springs can be geographically divided into three areas: the Rogers/Blue Point group (consisting of Springs 8 through 14 and numerous small springs and seeps); the Valley of Fire Wash group (Springs 5, 6, and 7); and Springs 2, 3, and 4 located further to the north. Many of these springs are related to regional structural features and generally have higher discharge rates and temperatures than locally-derived springs. Furthermore, these springs have similar isotopic compositions that are distinct from the compositions of the local springs.

Springs comprising the Rogers/Blue Point group are directly related to the Rogers Spring Fault, a major strike-slip fault in the Lake Mead area. The fault separates lower Paleozoic carbonate rocks of the Muddy Mountains on the northwest from Quaternary and Tertiary basin-fill deposits on the southeast. The low permeability basin-fill deposits form a barrier to eastward groundwater flow and cause the Rogers Spring Fault to act as a conduit [illegible] flow from the carbonates. Springs 8, 11, 12, and 13 issue directly from the fault, and Springs [illegible] 14 issue from the basin fill between the fault and Lake Mead. In addition, Spring 8 is located at the point of intersection of the Rogers Spring Fault and the Arrowhead Fault. Discharge rates of 1040 and 2750 L/min from Springs 8 and 11 (respectively) are the highest in the Lake Mead basin, reflecting the role of the Rogers Spring fault as an important conduit for groundwater flow in the region.

The regional nature of these springs is also reflected in the absence of a relation between discharge and precipitation patterns. Continuous measurements of the discharge rate at Spring 11 have been collected by the U.S. Geological Survey since October 1985. A comparison of the monthly discharge at Spring 11 (U.S. Geological Survey Water-Data Reports, Water Years 1984 through 1996) and the monthly precipitation in southern Nevada (based on data from 16 low elevation stations) (National Climatic Data Center, 1997) is shown in Figure 14. There is no consistent relationship between

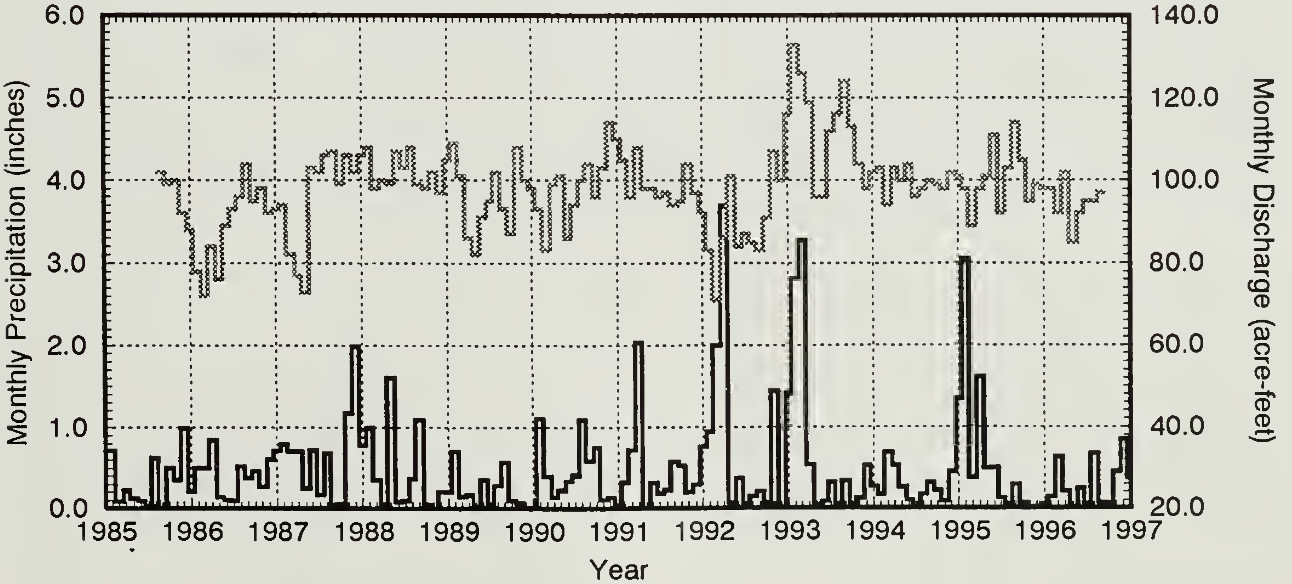

Figure 14. Comparison of monthly discharge at Spring 11 —— with monthly precipitation —— in southern Nevada.

precipitation and discharge, and although precipitation is generally greatest in the winter months when groundwater recharge is expected to be greatest, there is no consistent seasonal variation in discharge rate. This evidence suggests that discharge patterns at the North Shore Springs are more strongly related to regional flow than to local groundwater recharge.

In addition to the direct discharge represented by the North Shore springs, diffuse groundwater discharge occurs by evaporation and transpiration in several areas between the Muddy Mountains and the Overton Arm. Salt crusts on the soil surface indicate that evaporation from the water table is occurring near spring orifices and along drainage channels. Transpiration is indicated by thick stands of tamarisk, mesquite, acacia, various grasses, and other vegetation surrounding spring orifices and lining drainage channels. The amount of groundwater discharged by evapotranspiration (ET) may be significant relative to surface discharge at the spring orifice. Investigation of the amount of ET in the area of each spring was beyond the scope of the present study, though there is literature that can provide insight into the magnitude of groundwater discharge by this mechanism (Ball *et al.*, 1994; Smith *et al.*, 1996).

Springs of the Valley of Fire Wash group (Springs 5, 6, and 7) do not issue directly from the Rogers Spring Fault. Instead, Springs 5 and 6 are located in the area of an unconformable contact of Jurassic and Triassic clastic rocks on the west with the Tertiary Muddy Creek Formation on the east, near the Rogers Spring Fault. The mechanism of discharge is similar in that the springs occur where eastward flowing groundwater meets a low-permeability barrier formed by the Muddy Creek Formation and is forced upward, possibly along fault planes, to discharge points at ground surface. Spring 7 issues from Quaternary Older Alluvium near an exposure of the Muddy Creek formation.

The other subregional springs (Springs 2 and 3) in the Lake Mead basin are also unrelated to major structural features. Springs 2 and 3 are located near the unconformable contact of the Tertiary Horse Spring Formation on the west with the Tertiary Muddy Springs Formation on the east. Both springs are located in wash channels that cut through Overton Ridge, at the lowest land-surface elevations just upgradient from the low-permeability barrier of the Muddy Creek Formation. Thus, if groundwater in the area is assumed to be moving generally northeast or east toward the Muddy River and Colorado River, then these springs discharge at the intersection of the water table with land surface. Spring 4 issues from a gypsum unit within the Muddy Creek Formation.

Most subregional springs in the Lake Mead basin are of the mixed cation-sulfate composition (as shown in Figure 11), which is typical of the regional groundwaters in southern Nevada discussed earlier. Exceptions are the mixed cation-bicarbonate compositions of Springs 2 and 3, which will be discussed below. The generally higher Na and K concentrations of the subregional springs distinguish them from the local springs. Despite this relationship, this pattern does not represent an evolutionary trend from local springs to subregional springs in the Lake Mead basin because groundwater flow paths do not exist between the areas of local and subregional springs.

Despite differences in major ion chemistries, subregional springs in the Lake Mead basin show remarkably similar stable isotopic compositions (Figure 12); with the exception of Spring 4, their δD compositions range from -93.5 to -88 per mil. The stable isotope values of Spring 4 are indicative

of evaporation. The loose and open structure of the gypsiferous soil in the vadose zone near the spring and the high potential for evaporation from the slow moving water at the orifice suggest that significant evaporation occurs at the spring discharge point. A line extending from the subregional group to the composition of Spring 4 has a slope of about 2.6, which is consistent with kinetic isotopic enrichment during evaporation under conditions of low humidity. However, because Spring 4 issues from gypsum deposits, there is the possibility of altering the groundwater's isotopic composition by exchange and/or mixing with gypsum hydration water. Under dry conditions, gypsum can conserve its primary isotopic composition, but the exchange process is relatively rapid under wet conditions (Sofer, 1978). The effect of hydration water on groundwater composition would be a shift toward a heavier isotopic composition, reflective of the evaporated condition of the water that precipitated the gypsum. Thus, mixing with hydration water could account for the enriched composition of Spring 4, but without data on the gypsum composition, this cannot be proved. Despite their enrichment, the general coincidence of the isotopic composition of Spring 4 with other area groundwaters suggests the influence of hydration water, if any, is minimal, and that Spring 4 is subregionally-derived rather than local.

The stable isotopic compositions of the North Shore springs are isotopically lighter than locally-derived springs sampled in the Lake Mead basin, but are heavier than the regional carbonate aquifer at the terminal end of the White River Flow System (Figure 12). It is unlikely that the composition at the North Shore springs results from mixing isotopically lighter groundwater from the White River system with local, isotopically heavier groundwater because the volume of local recharge appears to be insufficient to cause the observed shift. A mixture of 75 percent groundwater having the composition of the Muddy River springs (average δD of -97.5 per mil) and 25 percent local recharge (average δD of -76 per mil) would be required to reach the composition of the North Shore springs. Twenty-five percent of the discharge of the North Shore springs is approximately 418 AFY (this value is a minimum since it does not include discharge by evapotranspiration), which is over 2.5 times larger than the amount of groundwater recharge estimated by Rush (1968) to originate from precipitation in the lower Moapa Valley, Black Mountains area (including the Muddy Mountains), and California Wash. In addition, extensive geologic evidence suggests that the Muddy River Springs form the terminus of the White River flow system (Dettinger *et al.*, 1995).

It is also unlikely that groundwater in the lower Virgin Valley is a major contributor to spring flow at the North Shore springs. Heads at the North Shore springs are higher than most of the heads measured by Metcalf (1995) in wells in the Older Alluvium in the Virgin River Valley, and higher than the altitude of the pre-Lake Mead confluence of the Muddy River and Virgin River, which lies between the Virgin Valley and the North Shore springs. Although limited to a single data point, the pre-Lake Mead hydraulic head near the confluence of the Muddy and Virgin Rivers appears to be approximately 265 m above mean sea level (Carpenter, 1915), which is 223 m below the Rogers Spring orifice. Furthermore, the Muddy Creek Formation may be more than 800 m thick below the Overton Arm and includes at least 300 m of very low permeability salt (Anderson and Laney, 1975). Finally, the limited volume of local, isotopically heavy groundwater is insufficient to cause the shift

from the very light groundwater in the Older Alluvium to the composition of the North Shore Springs.

The isotopic composition of the North Shore springs is in fact very similar to basin-fill aquifers in Weiser Wash, which appear to represent a mixture of groundwater moving south from Meadow Valley with groundwater recharged in the Mormon Mountains. This groundwater is isotopically heavier than the regional carbonate aquifer because these aquifers receive recharge from precipitation at lower elevations. Not surprisingly, the range of $\delta^{13}C$ values at Springs 8 and 11 (-3.9 to -6.2 per mil; Thomas *et al.*, 1991; Hershey and Mizell, 1995) indicate interaction with carbonate rocks, since these springs issue from carbonates. The ^{14}C values range from 3.0 to 7.2 PMC, indicating a long residence time in the groundwater system and the contribution of dead carbon from rock dissolution (uncorrected ages of approximately 20,000 to 30,000 years). The absence of atmospheric tritium in any North Shore springs indicates that all the groundwater is of a pre-1952 age.

Further discussion of the springs in Magnesite Wash and Kaolin Wash (Springs 2 and 3, respectively) is necessary here. These springs are located in wash channels that cut through Overton Ridge, down-gradient from a basin in Valley of Fire State Park that is comprised of Mesozoic sandstones and covered by thick, sandy soils. The lack of vegetation in this basin suggests that precipitation may infiltrate rapidly and is not available to support plant growth. The relatively low TDS contents of these springs (462 and 626 mg/L, respectively) suggest that they may originate from local recharge with minimal chemical interaction with the aquifer matrix in the basin, which is typical of groundwater flow in quartz arenites. However, the stable isotopic composition of these springs is much lighter than local, low-elevation recharge, instead plotting with the springs in the North Shore Complex. The $\delta^{13}C$ composition of these springs (-5.0 and -6.5) falls within the range of the North Shore Complex and indicates a contribution from dissolved carbonate minerals. Furthermore, the lack of atmospheric tritium indicates the groundwater residence time is relatively long. The apparent disagreement between the local origin suggested by the geographic and geochemical evidence and the subregional origin suggested by the isotopic evidence illustrates the complex hydrogeologic setting of these springs and indicates that their origin remains uncertain. However, one possible explanation is that these springs represent discharge from a subregional system that originates in the Mormon Mountains, as discussed below.

Taken as a whole, the isotopic data suggest that groundwater discharged at the North Shore Spring Complex is recharged in the region surrounding Lake Mead and is not directly related to flow in the regional carbonate aquifer of the White River Flow System. The most likely possibilities include the Muddy Mountains and the Mormon Mountains. Recharge in the Muddy Mountains alone is insufficient to provide the volume of discharge at the North Shore Springs. Evidence indicates that recharge in the Mormon Mountains represents the most likely source for the subregional flow system that discharges at the North Shore Spring Complex. Autochthonous Paleozoic carbonate rocks, well exposed throughout the mountains, provide the point of infiltration and recharge to the carbonate aquifer system. These autochthonous carbonate rocks continue southwest and plunge below ground surface at the Muddy Mountains. The autochthonous carbonate rocks are also exposed in the North

Muddy Mountains, though at lower elevations than at the Mormon Mountains. Not until crossing the Arrowhead fault do the autochthonous carbonate rocks descend completely into the subsurface, covered by the Mesozoic clastic formations and the allochthonous Paleozoic carbonate rocks of the Muddy Mountain thrust system. The autochthonous carbonate section is exposed again south of White Basin in the ridges just north of the Black Mountains. Here, the units are topographically much higher than at the major spring discharge of the subregional system at the Rogers/Blue Point complex.

The only structural obstruction in this flow path might occur near Glendale, just north of the North Muddy Mountains. It has been postulated that a strike-slip fault, the Moapa shear zone, separates the Mormon Mountains from the Virgin River depression to the south (Wernicke *et al.*, 1988). Whereas a major fault does separate the Mormon Mountains from the Tertiary sediments of the Virgin River depression, Anderson and Bernhard (1993) argue against a major through-going fault separating the North Muddy Mountains from the Mormon Mountains. The existence of this flow path is supported by evidence of groundwater discharge to the Muddy River reported by Rush (1968) in the reach passing through The Narrows at the northern edge of the North Muddy Mountains. This discharge indicates the presence of significant flow through the carbonate rocks between the Mormon and North Muddy Mountains, with upward flow occurring at favorable locations where overlying rocks are thin. Further evidence of this flow path may be provided by springs in Overton Ridge (Springs 2 and 3), that are located between The Narrows and the North Shore springs, are slightly lower in elevation than The Narrows, and have stable isotopic compositions indicative of subregional flow. Finally, the consistency of stable isotopic signatures of groundwater in the Mormon Mountains, Weiser Wash, Overton Ridge, and the North Shore Spring Complex indicate no major structural obstruction of the groundwater system's flow path until its primary discharge at the Rogers Spring Fault.

<u>Black Canyon</u>

In Black Canyon, Springs 26, 28, 29, and 30 are classified as subregional springs. Though these springs have widely varying temperatures (13° to 55°C) and discharge rates (13.2 to 960 L/min), their stable isotopic compositions are similar (as shown in Figure 13) and indicative of a common origin. In addition, these springs all possess a similar sodium and potassium-chloride composition (Figure 15), suggesting that their flow passes through rocks of similar minerology. Springs 26 and 30 issue from Tertiary volcanic rocks near northwest trending, right lateral strike-slip faults. Springs 28 and 29 issue from the Miocene Boulder City pluton at points where near vertical, north-south-trending faults intersect from below an unconformable barrier. This unconformity appears to act as a "ceiling", preventing further upward flow within the plutonic rocks.

The stable isotopic composition of Springs 26, 28, 29, and 30 is approximately midway between the end member compositions of subregional groundwater in Eldorado Valley, and local, low-elevation recharge. Note that using the Eldorado Valley water as an end-member is highly uncertain for the following reasons: only one sample is available from this basin; there are few data available from other, nearby deep basins; and there are no data from Arizona. Though Lake Mead

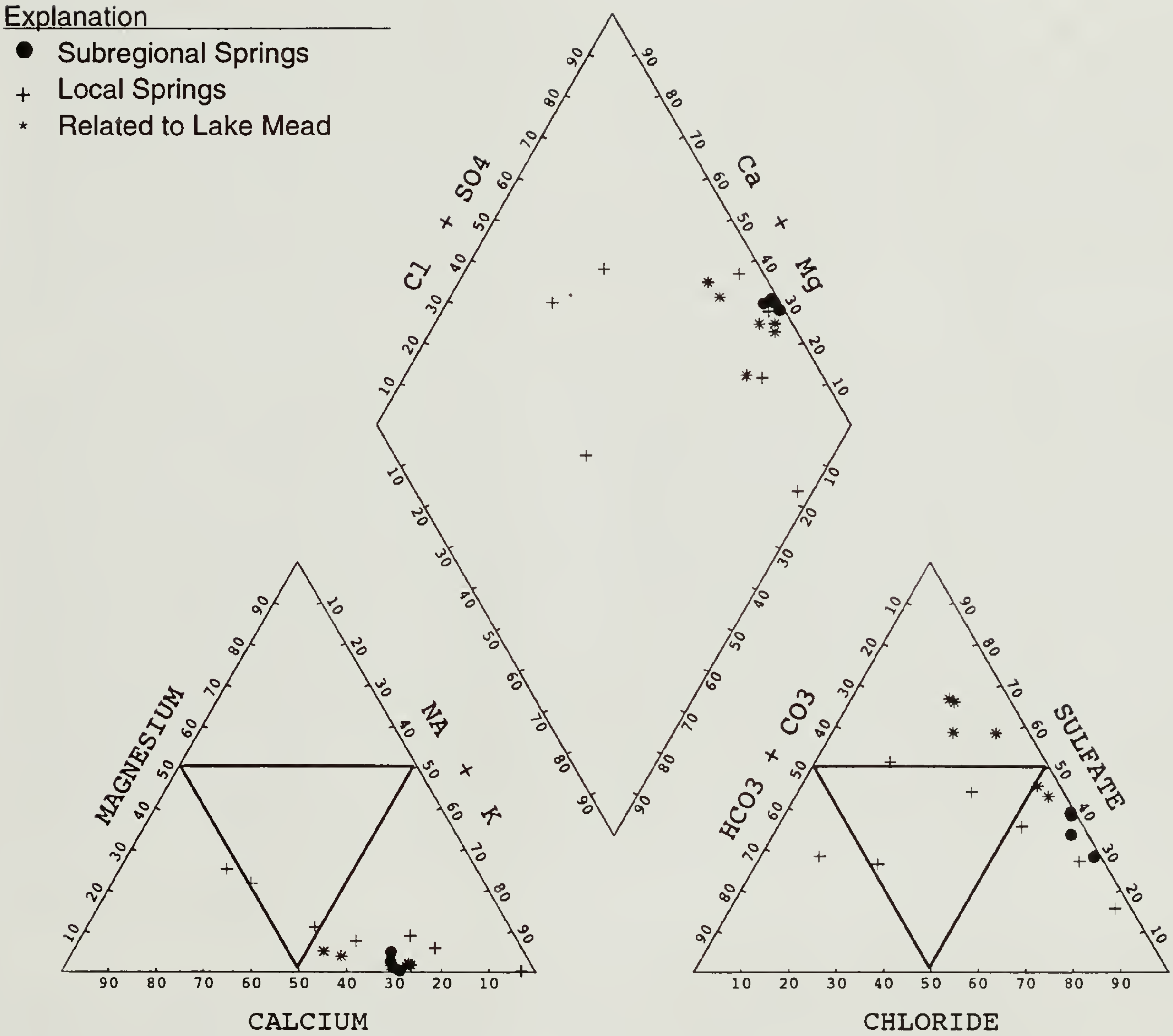

Figure 15. Trilinear diagram showing major dissolved ions of springs in the Black Canyon area.

water also represents a possible stable isotope end-member, Springs 26, 28, 29, and 30 can be distinguished from the springs affected by Lake Mead water by the following: with the exception of Spring 26, they contain no atmospheric tritium (the tritium content of Spring 26 is 21 pCi/L); they have TDS concentrations over 2000 mg/L; and they are at least 10 per mil enriched in δD with respect to springs located near the dam. Therefore, it appears unlikely that these springs are influenced by groundwater originating from Lake Mead.

Springs Influenced by Lake Mead Water

McKay and Zimmerman (1983) use environmental isotopes and water chemistry to demonstrate the hydraulic connection between Lake Mead and thermal springs in Black Canyon. Additional data collected for the present study confirm many of those results, and provide for some further refinement. Springs near Hoover Dam (Springs 20, 21, 22, 23, 24, and 25) share several

physical, geochemical, and isotopic properties: They tend to have the highest discharge rates and the highest temperatures (32° to 58°C) of springs in Black Canyon. Additionally, the TDS contents more closely resemble Colorado River water than other high discharge, subregional springs. The high discharge rates of many of the Black Canyon springs appear to result from the large hydraulic gradient imposed on the system by the altitude of the surface of Lake Mead, which is approximately 166 m above the river. The high temperatures reflect circulation near the Boulder City pluton. The temperature of Spring 20 is significantly lower than the others. This spring is closest to the dam, and the lower temperature may reflect less contact with the pluton than the other springs.

Springs near Hoover Dam also have the highest tritium activities (72 to 148 pCi/L) and the lightest $\delta^{18}O$ and δD values (δD of -106 to -100 per mil) (Figure 13). The high tritium activities indicate post-1952 groundwater recharge (a sample from the Colorado River on February 11, 1997 had a tritium activity of 51 pCi/L). The stable isotopes reflect the influence of Lake Mead water (a sample from the Colorado River on February 11, 1997 water had a δD content of -102 per mil). McKay and Zimmerman postulate a decreasing influence of the lake downstream, although they state that it is likely that all the springs in Black Canyon are influenced to some degree by Lake Mead. However, the tritium and stable isotope data collected during the present study suggest that the influence of Lake Mead water appears to end at a distance beyond Spring 25, which is 2.4 km downstream from the dam (Figure 16). Lake Mead water does not appear to impact Spring 26, which is within several hundred meters of Spring 25, is 35°C cooler, and is much more isotopically enriched. This suggests very different flow paths and/or origins for these two springs. Spring 26 is considered a subregional spring, as discussed above.

Uranium Signatures

The uranium data gathered for this study are shown, along with pertinent data from other sources, in Figure 17. The springs shown in this plot can be divided into two major groups – one with high uranium concentrations and low activity ratios (Springs 4, 15, 16, 17, 18, 19, and 36), the other with higher ARs, but generally lower uranium concentrations than the first group (Springs 1, 2, 3, 6, 7, and 11). The uranium signature of the first group suggests residence times which are relatively short, as relatively little leaching has taken place. The second group appears to have had a longer residence time, as increased leaching has caused a shift in the U signatures to a higher AR, with lower concentration. One obvious explanation for the different uranium signatures relates to the source area for any given spring – water discharging from springs which have a local source would have relatively short flowpaths, while water discharging from springs which have source areas outside local basins would typically require a longer transport time from recharge to discharge. Thus, locally-derived springs would display low activity ratios and high concentrations, and regional springs would display high ARs and low concentrations.

The springs that exhibit high concentrations and low ARs share similar uranium isotope signatures with locally-derived springs in the Virgin Mountains (the lower most triangles in Figure 17). With the exception of Spring 4, the uranium isotope signature of these springs supports their geographic and stable isotope designation as local springs. The stable isotope data suggests Spring 4

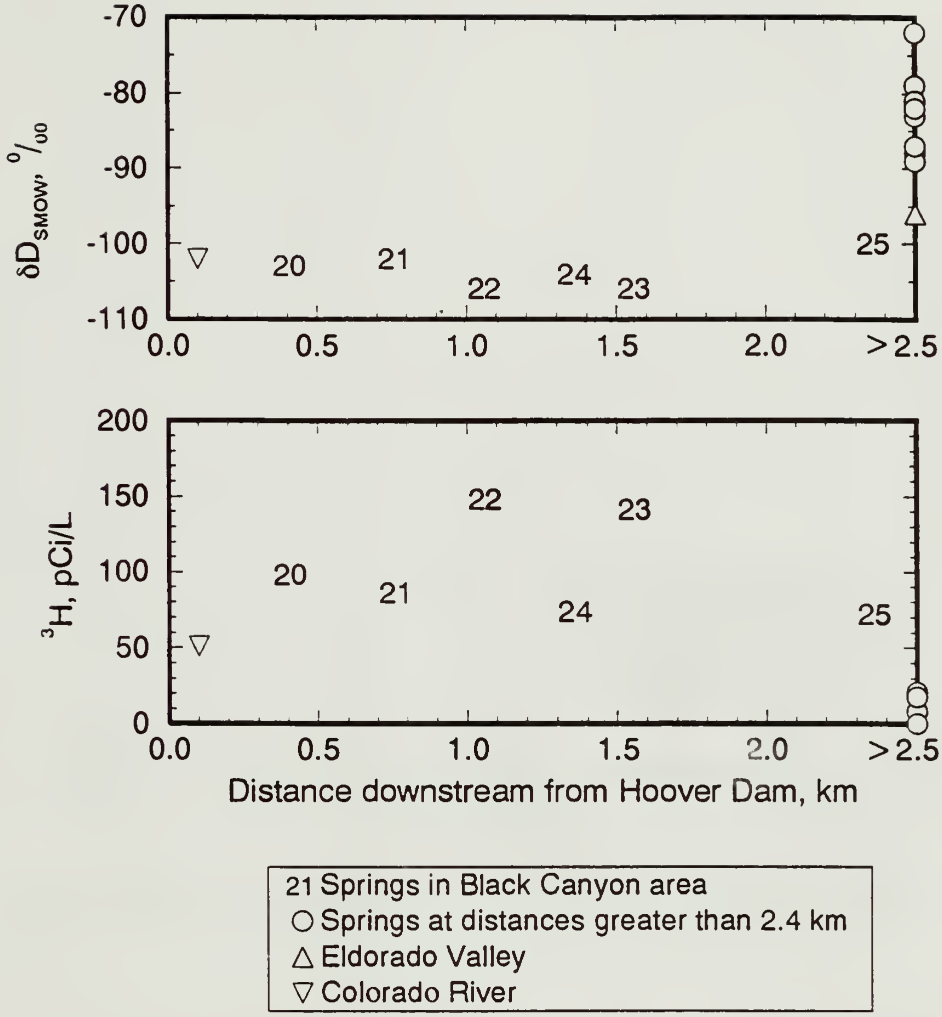

Figure 16. Plot of δD and ^3H as a function of distance downstream from Hoover Dam.

is derived from flow outside the local basin and that the discharge has been subjected to evaporation, as discussed above. The uranium isotope signature of the other group of springs is indicative of longer residence times, and supports their designation as subregional springs based on geographic and geologic settings and the stable isotopic data. For the most part, these springs have lower ARs than other regional springs in southern Nevada and southwestern Utah for which data are available. Although regional data for uranium are not as abundant as data for stable isotopes, the recent studies by Farmer (1996) and Yelken (1996) may be indicative of broader awareness and acceptance of uranium-series disequilibrium as an interpretive tool for investigating groundwater flow in southern Nevada. If this is the case, further interpretation of spring sources and water evolution along flowpaths will be possible as the regional uranium database grows.

For springs in the Valley of Fire Wash group, the uranium data may provide additional insight into flow patterns delineated using stable isotope data. Stable isotope data in non-geothermal systems provides information on initial recharge conditions and any subsequent

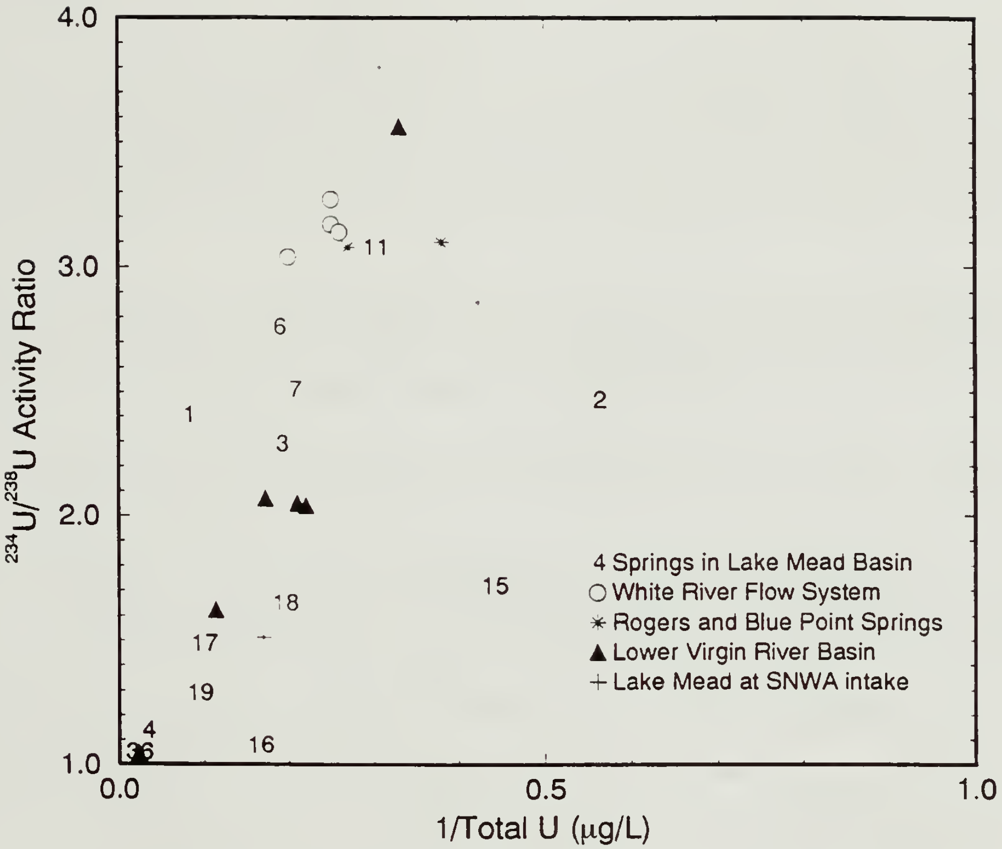

Figure 17. Uranium composition of springs in the Lake Mead basin, as compared to other waters in the region.

evaporation, but typically are not affected by water-rock interaction. Uranium isotope values can change as water moves along a flowpath and evolves due to interaction with aquifer materials.

Springs 6 and 7, which plot together with the Rogers/Blue Point group on the stable isotope graph, exhibit lower activity ratios and slightly higher concentrations than do Rogers and Blue Point springs. This may be suggestive of mixing between a lower-concentration, higher-AR water (discharge at Rogers and Blue Point springs) and a water which is leaching a "fresher" source of uranium. Water flowing through a rock body which has had less leaching take place would tend to provide a higher uranium concentration, but a lower AR than a water interacting with a more highly leached rock body (Osmond and Cowart, 1992). Perhaps, then, the springs in Valley of Fire Wash share a common water source with Rogers and Blue Point Springs, but are more recent in origin.

The uranium signature for Spring 2 indicates a significantly lower concentration than the North Shore Complex springs, to which it is related by location and stable isotope signature. Since evaporation is not apparent in the stable isotope data, the two most likely explanations for the uranium signature are either dilution at some point along the flowpath for this spring, or flow toward other springs in the group passing through a localized area of anomalously high uranium concentration.

CONCLUSIONS

Thirty-six springs in and around the Overton Arm, Boulder Basin, and Black Canyon areas of the Lake Mead National Recreation Area were visited and described. Historical data, which generally included discharge measurements, chemical indicator measurements, and water chemistry analyses, were compiled and supplemented by stable and radioactive isotopic data collected during the present study.

Three classifications of source area have been defined for the springs, primarily based on hydrogeologic setting and the stable isotopic data. Distinguishing characteristics of these three classifications, and the springs included in each, are listed in Table 3.

Table 3. Characteristics of the Three Spring Classifications Defined by this Study, and the Springs Included in Each.

	Local	Subregional	Lake Mead
Geologic Setting	Generally not related to regional structural features.	Often related to regional structural features such as faults.	Related to normal faulting around Boulder City pluton.
Discharge Rate	Less than 20 L/min, most less than 3 L/min	1 to 2750 L/min	10 to 1540 L/min
Temperature	10 to 25°C	15 to 30°C	32 to 58°C
δD	-67 to -80 per mil	-88 to -93 per mil	-106 to -100 per mil
^3H	Less than 5 to 18 pCi/L	Less than 5 pCi/L	74 to 141 pCi/L
Uranium Activity Ratio	Less than 2.0	Greater than 2.0	—
Spring Name and ID	Kelsey (1) Bitter (15) Sandstone (16) Cottonwood (17) Gypsum (18) Unnamed, in Rainbow Gardens (19) Unnamed, in Horsethief Canyon (27) Unnamed, near Spring 30 (31) Nevada Falls (32) Bighorn Sheep (33) Arizona Seep (34) Latos Pool (35) Unnamed, in Aztec Wash (36)	Unnamed, in Magnesite Wash (2) Unnamed, in Kaolin Wash (3) Getchel (4) Unnamed, in Valley of Fire Wash (5) Unnamed, in Valley of Fire Wash (6) Unnamed, in Valley of Fire Wash (7) Blue Point (8) Unnamed (9) Unnamed (10) Rogers (11) Scirpus (12) Corral (13) Unnamed (14) Palm Tree, Cold (26) Boy Scout Canyon, Hot (28) Boy Scout Canyon, Cold (29) Arizona Hot Spring (30)	Pupfish (20) Arizona Hot Spot (21) Sauna Cave (22) Nevada Hot Spring (23) Nevada Hot Spot (24) Palm Tree, Hot (25)

Almost one third of the springs studied are considered to be of local origin. Locally-derived springs discharge groundwater from small flow systems that receive most or all of their recharge

locally and at low altitudes. These springs are generally not related to major structural features, instead discharging from small fractures or joints, or the bottoms of wash channels. The low discharge rates of local springs result from the limited groundwater recharge that occurs at low elevations in this arid region. Temperatures are lower than the other springs because of rapid equilibration of the low volume discharge with ambient land surface and air temperatures, and because groundwater does not circulate to great depths. The stable isotopic values are indicative of low-elevation recharge in southern Nevada. Low uranium activity ratios and relatively higher uranium concentrations are indicative of relatively short residence times, which generally result from shorter flow paths, and support the designation of these springs as locally derived. Despite their local origin, however, non-detectable to very low tritium concentrations suggest travel times longer than several decades and very limited recharge by recent precipitation events.

Local springs are unrelated to regional groundwater flow systems such as the carbonate aquifer system. For springs in the Lake Mead basin, recharge occurs in the Black Mountains, Bitter Spring Valley (and possibly the slopes of surrounding ridges), and the area surrounding Rainbow Gardens. Local springs in Black Canyon originate from recharge in the Black Mountains and Eldorado Mountains. Most of the local springs in the recreation area discharge from localized groundwater flow systems that are contained within the park boundaries. Although the Maxey-Eakin method predicts that groundwater recharge is neglibible at low elevations in southern Nevada, the existence of these springs indicates that certain geologic, topographic, climatic, and hydrologic conditions can combine to produce local flow systems that are capable of supplying perennial springs. The small sizes of these flow systems, which suggests that their groundwater storage potential is small, means that locally-derived springs are more sensitive to local climate and recharge conditions than the larger, subregional springs, and therefore may require special management and protection.

Subregional springs are dominated by groundwater that originates outside local flow systems, and therefore outside the recreation area, and may include groundwater recharged at higher elevations. The locations of subregional springs are often related to major, regional structural features. Most of the subregional springs in the Lake Mead basin (the Rogers/Blue Point and Valley of Fire Wash groups) are related to the Lake Mead strike/slip fault system, while most of the subregional springs in the Black Canyon area are related to a system of north-south-trending normal faults. Most of these springs represent the ultimate discharge of subregional groundwater flow systems and therefore have higher discharge rates than the local springs. Their higher temperatures result from deeper circulation and less equilibrium with ambient land surface and air temperatures. The stable isotopic values are indicative of higher elevation recharge sources than most of the region surrounding Lake Mead. Non-detectable tritium concentrations and low percentages of modern carbon indicate that these waters have long residence times. Higher uranium activity ratios are indicative of longer residence times, and generally longer groundwater flow paths, where the water has more time in contact with the rock.

Subregional springs in the Lake Mead basin appear to be most strongly related to groundwater systems that extend north to the Weiser Wash and Mormon Mountains area, rather than to the regional White River Flow System or Virgin River basin. Subregional springs in the Black Canyon

area appear to originate from a mixture of subregional flow (e.g., Eldorado Valley in Nevada, possibly Detrital Valley in Arizona) and local, low-elevation recharge in the Black Mountains and Eldorado Mountains. The subregional origin of these springs suggests that they may be more sensitive than previously thought to groundwater impacts in the areas adjacent to the park.

A third set of springs is derived from recirculated Lake Mead water, as first described by McKay and Zimmerman (1983). These springs are related to normal faulting around the Boulder City pluton, which provides the heat source for their high temperatures. The high discharge rates exhibited by several of these springs probably relate to the very high gradient of hydraulic head that results from the impoundment of Lake Mead by Hoover Dam. The stable isotope values form a range around the present composition of the Colorado River, implicating it as the most probable source. In addition, the tritium contents of these springs indicates that at least a portion of these waters were recharged after 1952.

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APPENDIX A

PHYSICAL, CHEMICAL, AND ISOTOPIC DATA

Table A-1. Field Measurements.

ID	Latitude (d m s)	Longitude (d m s)	Altitude (m)	Discharge Rate (L/min)	Temp (°C)	EC (μS/cm)	pH (Std. Units)	DO (mg/L)	HCO_3 (mg/L)	Date
1	36 31 38	114 24 51	375	<1	22	3561	7.05	3.8	147	03/07/96
2	36 29 59	114 28 35	427	<1	16	470	7.85	8.1	—	10/04/95
2	36 29 59	114 28 35	427	<1	11.1	552	8.25	8.8	200	02/09/96
3	36 29 14	114 28 00	439	<1	19	545	8.13	7.6	—	10/04/95
3	36 29 14	114 28 00	439	<1	14.1	770	8.46	6	180	02/09/96
4	36 26 36	114 24 17	424	<1	10.8	23905	7.88	8.8	—	02/09/96
5	36 24 21	114 26 38	450	~1	15	3590	7.61	5.25	156	03/07/06
6	36 24 19	114 25 50	450	13.1	13.5	8024	7.76	3.9	118	03/07/96
7	36 24 05	114 24 07	381	~40	23	5520	7.1	5	—	02/09/96
8	36 23 24	114 25 59	470	—	30	4535	7.03	2.1	—	10/04/95
8	36 23 24	114 25 59	470	1040	29.6	4270	7.05	2.65	—	02/08/96
9	36 22 59	114 26 00	494	<1	17	4235	8.02	7	—	02/08/96
10	36 22 45	114 25 30	430	>40	15	8100	7.55	7.5	—	02/08/96
11	36 22 37	114 26 40	488	2750[2]	30	4190	7.22	4.6	—	10/03/95
11	36 22 37	114 26 40	488	—	30	3860	7.03	2.6	—	02/08/96
12	36 22 37	114 26 57	480	<1	17	4935	7.13	0.7	—	02/07/96
13	36 22 14	114 27 36	485	<1	17	4315	7.31	6.2	152	02/07/96
14	36 21 28	114 26 14	396	30	17	5590	8.04	8.6	—	02/08/96
15	36 17 06	114 30 51	506	12	25	4090	7.43	3.15	—	10/03/95
15	36 17 06	114 30 51	506	—	17.2	4021	7.58	4.75	104	02/06/96
16	36 12 40	114 33 24	601	<1	19	1265	7.06	1.25	—	10/03/95
16	36 12 40	114 33 24	601	<1	11	1450	7.03	1.95	146	02/07/96
17	36 12 12	114 38 37	661	<1	18	3690	7.63	2.4	—	10/03/95
17	36 12 12	114 38 37	661	0.07	12.6	3625	7.81	6.5	173	02/06/96
18	36 12 29	114 54 44	530	<1	22	4860	7.56	7.2	—	10/02/95
18	36 12 29	114 54 44	530	<1	15.8	4230	7.38	4.2	114	02/06/96
19	36 06 26	114 58 10	500	<1	25	4900	7.05	2.5	—	10/02/95
19	36 06 26	114 58 10	500	<1	15.5	4785	7.81	3.8	129	02/05/96
20	36 00 40	114 44 35	240	636	36	1204	7.79	3.3	—	02/11/97
21	36 00 05	114 44 30	210	60	55.1	2775	7.62	3.1	—	01/31/97
22	36 00 11	114 44 36	220	22.2	45	1893	7.66	4	—	02/01/97
23	36 00 10	114 44 58	280	1536	46	1788	7.36	1.6	—	01/31/97
24	36 00 04	114 44 36	210	18	58	2323	8	3	—	01/31/97
25	35 59 43	114 44 19	230	10.2	48	3599	7.55	2.5	—	02/01/97
26	35 59 41	114 44 15	235	13.2	13	7059	7.95	10.0	—	02/01/97
27	35 59 56	114 37 58	988	2	12	1069	7.66	—	—	02/03/97
28	35 58 59	114 44 49	260	960[1]	55	4601	7.43	1.9	—	02/02/97
29	35 58 59	114 44 49	263	—	24	4313	7.10	8.0	—	02/02/97
30	35 57 39	114 43 32	245	126	44	4991	7.70	2.4	—	02/01/97
31	35 57 39	114 43 32	249	4.2	19	3368	7.78	6.8	—	02/01/97
32	35 56 43	114 43 55	211	8.4	19	1022	7.34	—	—	02/02/97
33	35 56 21	114 44 03	245	10.2	32	816	7.92	4.2	—	02/02/97
34	35 55 35	114 42 24	220	<1	24	7171	7.47	—	—	02/03/97
35	35 50 55	114 43 33	293	2	25	750	8.08	4.5	—	05/06/97
36	35 39 36	114 46 20	605	<1	18	1505	7.34	1.7	—	02/05/96
36	35 39 36	114 46 20	605	2	15	1874	7.54	3.45	—	02/11/97
ES[3]	35 48 13	115 00 14	550	—	23	891	8.68	3.6	96	05/02/97
CR[3]	36 00 35	114 44 40	200	—	14	927	8.18	8.4	—	02/11/97

[1] Combined discharge of hot and cold springs
[2] Annual mean based on water years 1985 to 1994 in U.S.G.S. Water-Data Reports
[3] ES Eldorado Substation Well
CR Colorado River, below Hoover Dam

Table A-2. Major Ion and Trace Metal Chemistry.

ID	EC (lab) (µS/cm)'	pH (lab) Std. Units	TDS mg/L	HCO_3 (lab, mg/L)	CO_3 mg/L	Cl mg/L	SO_4 mg/L	NO_3 mg/L	Na mg/L	K mg/L	Ca mg/L	Mg mg/L	SiO_2 mg/L	Li mg/L	F mg/L	Fe mg/L	N mg/L	B mg/L	Mn mg/L	Date	Source[1]
1	3640	7.5	2721	319	—	332	1180	—	547	25.4	156	92.9	46.8	—	—	—	0.84	—	—	03/07/96	f
2	574	8.27	462	249	0	18.9	60	—	52	25.6	34.7	16.8	5	—	—	—	0.04	—	—	02/09/96	f
3	819	8.35	626	213	2.2	46.5	168	—	77.6	21.3	48.9	25.9	19.1	—	—	—	3.32	—	—	02/09/96	f
4	17500	8.1	16300	270	0	2100	8800	—	3800	300	470	610	54	—	2.6	0.2	—	17	0.3	05/17/78	g
5	4350	7.62	3841	169	—	278	2290	—	295	51.1	537	208	12.4	—	—	—	—	—	—	03/07/06	f
6	—	7.9	9970	240	0	1900	4800	—	1900	130	590	510	16	1.1	2	0.01	0.05	7.3	0.02	05/05/77	g
7	—	—	4710	140	0	600	2600	—	600	45	510	260	24	—	1.6	0	—	2.2	0.03	05/19/78	g
8	4100	7.8	—	160	—	400	1900	—	330	23	470	160	16	0.68	1.5	<0.009	0.2	—	<0.003	07/01/85	a
9	4190	7.5	3710	220	0	370	2100	—	340	27	560	180	21	—	1.5	0.01	—	1.4	0.01	05/19/78	g
10	—	7.7	9270	650	0	1700	4300	—	1700	130	300	720	89	—	4.5	0.02	—	6.2	0.04	05/19/78	g
11	—	7.48	—	161	0	327	1620	—	291	22.7	423	143	16.8	—	1.4	—	0.27	—	—	03/19/92	e
12	4440	7.6	3787	266	0	386	2040	—	350	25.3	513	186	20.4	—	—	—	<0.04	—	—	02/07/96	f
13	—	—	3440	180	0	400	1900	—	340	23	510	160	16	0.7	1.6	0.01	0.1	1.3	0.01	05/04/77	g
14	5600	—	4930	170	0	680	2700	—	580	18	580	250	33	0.96	2	0.03	0	1.8	0.02	05/19/77	g
15	4200	8.1	3730	140	0	160	2400	—	270	22	580	190	33	0.83	2.8	0.01	0.02	1.5	0.01	05/03/77	g
16	1550	7.58	1215	249	0	16.9	725	—	21.9	4.96	209	79.2	13.8	—	—	—	1.15	—	—	02.07/96	f
17	3890	7.99	3660	205	0	63.6	2410	—	209	10.7	524	220	17.4	—	—	—	<0.04	—	—	02/06/96	f
18	4450	7.79	4253	146	0	151	2840	—	231	21.5	532	308	23.6	—	—	—	<0.04	—	—	02/06/96	f
19	5280	7.62	4931	144	0	379	3040	—	405	38.8	569	332	13.9	—	—	—	9.26	—	—	02/05/96	f
20	1250	—	757	116	—	108	335	1.77	188	4.5	59.9	2.9	29.6	—	—	—	0.4	—	—	05/02/95	b
21	—	—	1749	77.5	—	476	589	<0.4	451	9.48	140	5.9	34.3	—	—	—	—	—	—	01/31/97	f
22	1780	—	1280	141	—	134	584	0.27	210	8.27	150	11	50.1	—	—	—	0.06	—	—	05/02/95	b
23	1780	—	1260	135	—	145	584	0.31	226	7.53	138	8.4	48	—	—	—	0.07	—	—	05/02/95	f
24	2340	—	1580	98.1	—	283	644	<0.04	343	8.24	133	4.1	48.7	—	—	—	<0.01	—	—	05/02/95	b
25	—	—	2017	70.7	—	591	644	0.09	522	11	172	6.11	36.8	—	—	—	—	—	—	02/01/97	f
26	—	—	4235	151	—	1514	1100	0.31	1030	16.9	382	40.7	53.6	—	—	—	—	—	—	02/01/97	f
27	—	—	784	324	—	101	140	<0.04	51.8	11.9	120	34.8	62	—	—	—	—	—	—	02/03/97	f
28	4500	—	2920	28.5	—	956	843	0.09	695	14.4	247	2.3	44.7	—	—	—	0.02	—	—	05/03/95	b
29	4500	—	2490	33.2	—	962	827	<0.04	680	15.8	257	6	51.7	—	—	—	<0.01	—	—	05/03/95	b
30	—	—	2635	37.3	—	1078	587	4.92	650	15.2	249	13.5	39.4	—	—	—	—	—	—	02/01/97	f

Table A-2. Major Ion and Trace Metal Chemistry (Continued).

ID	EC (lab) (μS/cm)	pH (lab) Std. Units	TDS mg/L	HCO_3 (lab, mg/L)	CO_3 mg/L	Cl mg/L	SO_4 mg/L	NO_3 mg/L	Na mg/L	K mg/L	Ca mg/L	Mg mg/L	SiO_2 mg/L	Li mg/L	F mg/L	Fe mg/L	N mg/L	B mg/L	Mn mg/L	Date	Source[1]
31	—	—	1933	97.9	—	755	410	11	407	11	212	28.6	33.2	—	—	—	—	—	—	02/01/97	f
32	1210	—	691	83.8	—	194	182	6.91	178	3.14	38.2	7.5	29	—	—	—	1.56	—	—	05/03/95	b
33	820	—	493	81.5	14.6	89.7	145	12.5	164	0.89	4.26	0.3	24.4	—	—	—	2.82	—	—	05/04/95	b
34	—	—	4228	148	—	2019	525	18.5	1103	25.4	313	76.4	40.6	—	—	—	—	—	—	02/03/97	f
35	347	—	215	112	6.7	13.5	42	8.95	34.3	6.46	28.5	4.6	11.1	—	—	—	2.02	—	—	04/20/95	b
36	—	—	1298	363	—	102	445	<0.04	118	5.89	178	47.6	38.9	—	—	—	—	—	—	02/05/96	f
ES[2]	795	—	498	124	10.6	119	60.9	14.4	154	4.01	6.84	0.5	64.9	—	—	—	3.24	—	—	09/26/95	b
CR[2]	—	—	689	160	—	80	245	1.51	94.6	4.87	74.4	28.3	8.7	—	—	—	—	—	—	02/11/97	f

[1] Sources of data:
a Thomas *et al.*, 1991
b SNWA, unpublished data
e Hershey and Mizell, 1995
f This study
g Laney and Bales, 1996

[2] ES Eldorado Substation Well
CR Colorado River, below Hoover Dam

Table A-3. Isotopic Compositions.

ID	^{3}H (pCi/L)	$\delta^{18}O$ (per mil)	δD (per mil)	$\delta^{13}C$ (per mil)	14 PMC	$^{234}U/^{238}U$ (act. ratio)	Total U (μg/L)	Date	Source[1]
1	—	-10	-82	-7.6	—	2.41	13.3	03/07/96	f
2	<10	-11.5	-92	-5.0	—	2.47	1.8	02/09/96	f
3	<10	-11.3	-88	-6.5	—	2.29	5.45	02/09/96	f
4	<10	-8.6	-83	—	—	1.14	37.9	02/09/96	f
5	—	-12.2	-93	—	—	—	—	03/07/96	f
6	—	-11.8	-92	—	—	2.76	5.50	03/07/96	f
7	—	-11.2	-88	-6.8	—	2.51	5.0	02/09/96	f
8	—	-12.4	-93	-6.2	3.5	3.07	—	06/24/85	a
8	—	-12.5	-93.5	-5.3	7.2	—	—	07/01/85	a
8	<10	-12.3	-91	—	—	—	—	02/08/96	f
11	<10	-12.4	-92	-3.9	3	~4.0	~2.9	03/19/92	e
11	—	-12.4	-91	—	—	3.08	3.49	02/08/96	f
12	—	-12	-90	—	—	—	—	02/07/96	f
13	—	-12.1	-91.5	—	—	—	—	02/07/96	f
15	<10	-9.9	-77	-4.3	—	1.72	2.35	02/06/96	f
16	—	-10.5	-79	—	—	1.08	6.69	02/07/96	f
17	<10	-10.8	-80	—	—	1.49	12.0	02/06/96	f
18	<10	-9.2	-75	—	—	1.65	5.59	02/06/96	f
19	<10	-8.6	-71	—	—	1.29	12.8	02/05/96	f
20	—	—	—	-6.6	—	—	—	02/11/97	f
20	98	-12.9	-103	—	—	—	—	05/02/95	b
21	86	-13	-102	-7.4	—	—	—	01/31/97	f
22	148	-13.7	-106	—	—	—	—	05/02/95	b
23	141	-13.6	-106	-28.65	62.9	—	—	05/02/95	b
24	74	-13.5	-104	—	—	—	—	05/02/95	b
25	72	-12.7	-100	-8.0	—	—	—	02/01/97	f
26	21	-11.2	-88	-11.8	—	—	—	02/01/97	f
27	8	-10.8	-79	—	—	—	—	02/03/97	f
28	<10	-11.5	-92	-27.64	26.98	—	—	05/03/95	b
29	<10	-10.8	-88	—	—	—	—	05/03/95	b
30	<5	-11.2	-87	-11.5	50.71	—	—	02/01/97	f
31	<5	-10.3	-81	-13.2	81.82	—	—	02/01/97	f
32	<10	-10.2	-83	—	—	—	—	05/03/95	b
33	<10	-10.3	-83	-24.91	15.34	—	—	05/04/95	b
34	<5	-10.3	-82	-7.0	—	—	—	02/03/97	f
35	8.2	-9.8	-81	-11.9	—	—	—	05/06/95	f
36	18	-9.2	-72	—	—	1.05	134	02/05/96	f
36	—	—	—	-13.2	—	—	—	02/11/97	f

[1] Sources of data:
a Thomas *et al.*, 1991
b SNWA, unpublished data
e Hershey and Mizell, 1995
f This study

APPENDIX B

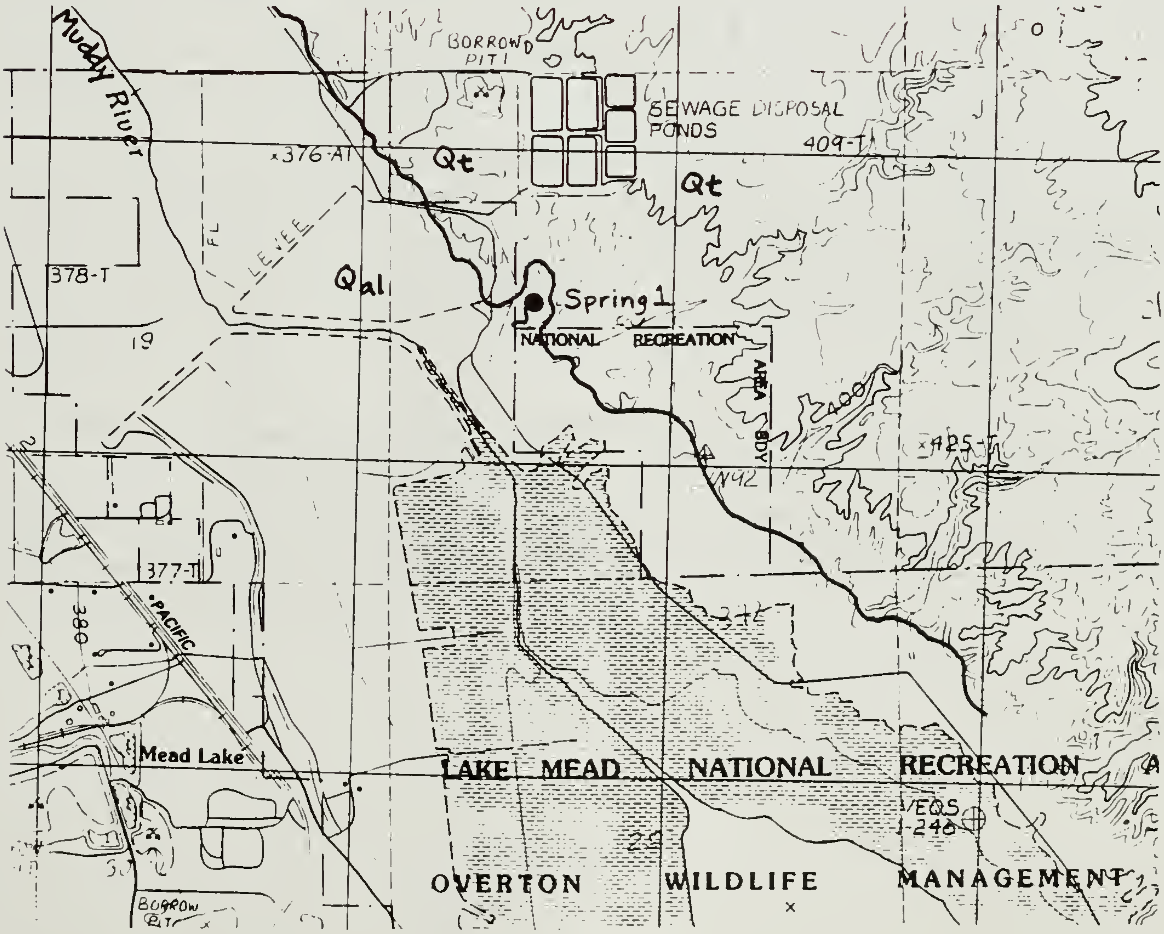
GEOLOGIC DESCRIPTIONS

Spring 1 – Kelsey Spring

Topographic base: 7.5' Overton Quadrangle
Geology references: This study

Kelsey Spring is located at the northeast edge of the Overton Wildlife Management area at the base of Mormon Mesa. The orifice is covered by a concrete vault (having dimensions 1.5 by 2 m wide and 1.5 m high) with an access door in the top. There was approximately 0.75 m of standing water in the vault when this spring was visited on 3-7-96. Seepage from the vault occurs in cracks in the concrete near its base. Samples were collected from this seepage. In addition, a 10-cm-diameter steel pipe extends south about 20 m from the vault and discharges at ground surface within a stand of very dense vegetation. A large area of reeds extends north and slightly uphill from the vault, suggesting that groundwater is near ground surface and that additional discharge may be occurring in that area.

Kelsey Spring discharges near the base of Quaternary terrace deposits at the edge of Mormon Mesa. Other seeps are located at the base of the terrace, as indicated by several stands of palm trees to the northwest.



Spring 2 – Unnamed spring in Magnesite Wash

Topographic base: 7.5' Overton Quadrangle
Geology references: Bohannon (1983)

The spring is located in a gap in Overton Ridge though which the Magnesite Wash channel passes. The spring issues as subsurface discharge into a 10-m-diameter pool. Additionally, minor seepage can be observed up to 5 m above the pool from fractures in the Tertiary Basal Conglomerate. Surface flow occurs for only a few 10s of meters downstream from the pool, which is surrounded by reeds, willows, and grape vines.

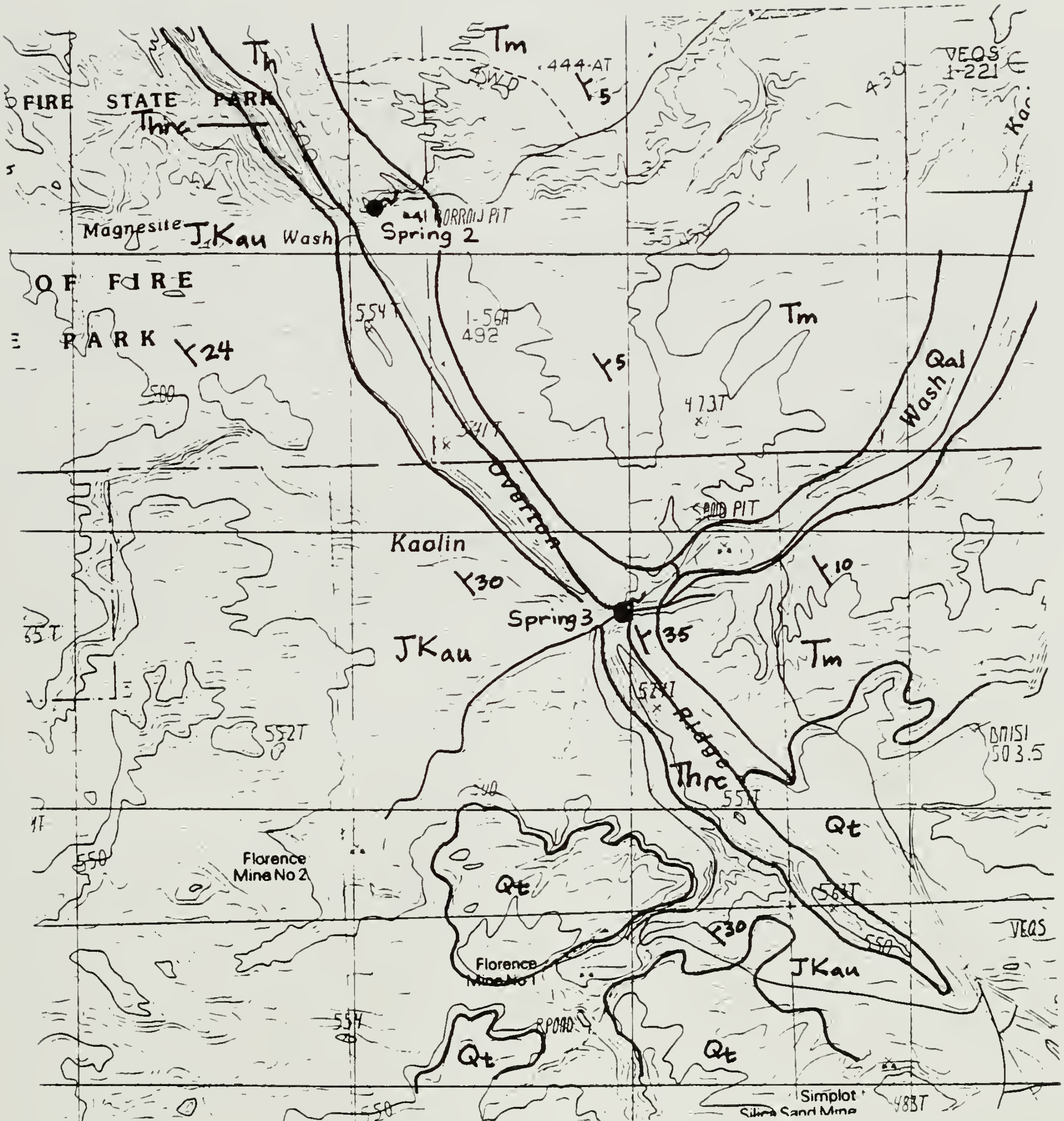
The spring is located at the contact of the Basal Conglomerate with the upper Rainbow Gardens Member (both of the Tertiary Horse Spring Formation), and about 200 m west of the unconformable boundary with the Tertiary Muddy Springs Formation. The spring is not associated with any major structural features. Rather, if groundwater is assumed to be moving generally west or northwest toward the Muddy River and Colorado River, then the spring is located at the lowest elevation just upgradient from the low-permeability barrier of the Muddy Creek Formation. Upstream of the spring, Magnesite Wash passes through a basin comprised of Mesozoic sandstones and covered by thick, sandy soils.

Spring 3 – Unnamed spring in Kaolin Wash

Topographic base: 7.5' Valley of Fire, East Quadrangle
Geology references: Bohannon (1983)

The setting for this spring is similar to the Magnesite Wash spring; a gap in Overton Ridge through which Kaolin Wash passes, although the gap at Kaolin Wash is much narrower. At Kaolin wash, the spring issues as subsurface discharge into a 5-m-diameter pool. Additionally, minor seepage can be observed from fractures in the Thumb Member. Surface flow occurs for approximately 400 m downstream from the pool, which is surrounded by reeds.

The spring issues from the Thumb Member of the Teriary Horse Spring Formation, and about 1 km upstream (southwest) of the contact between the the Thumb Member and the Muddy Creek Formation. And, similar to the Magnesite Wash spring, the Kaolin Wash spring is located near the lowest elevation just upgradient from the low-permeability barrier of the Muddy Creek Formation.


Tm
Th
FIRE STATE PARK
Thrc
Magnesite
JKau
Wash
Spring 2
BORROW PIT
OF FIRE
PARK
Tm
Qal
Wash
Overton
Kaolin
SAND PIT
Spring 3
JKau
Tm
Ridge
Thrc
Qt
Florence
Mine No 2
Qt
Florence
Mine No 1
JKau
Qt
Qt
Simplot
Silica Sand Mine

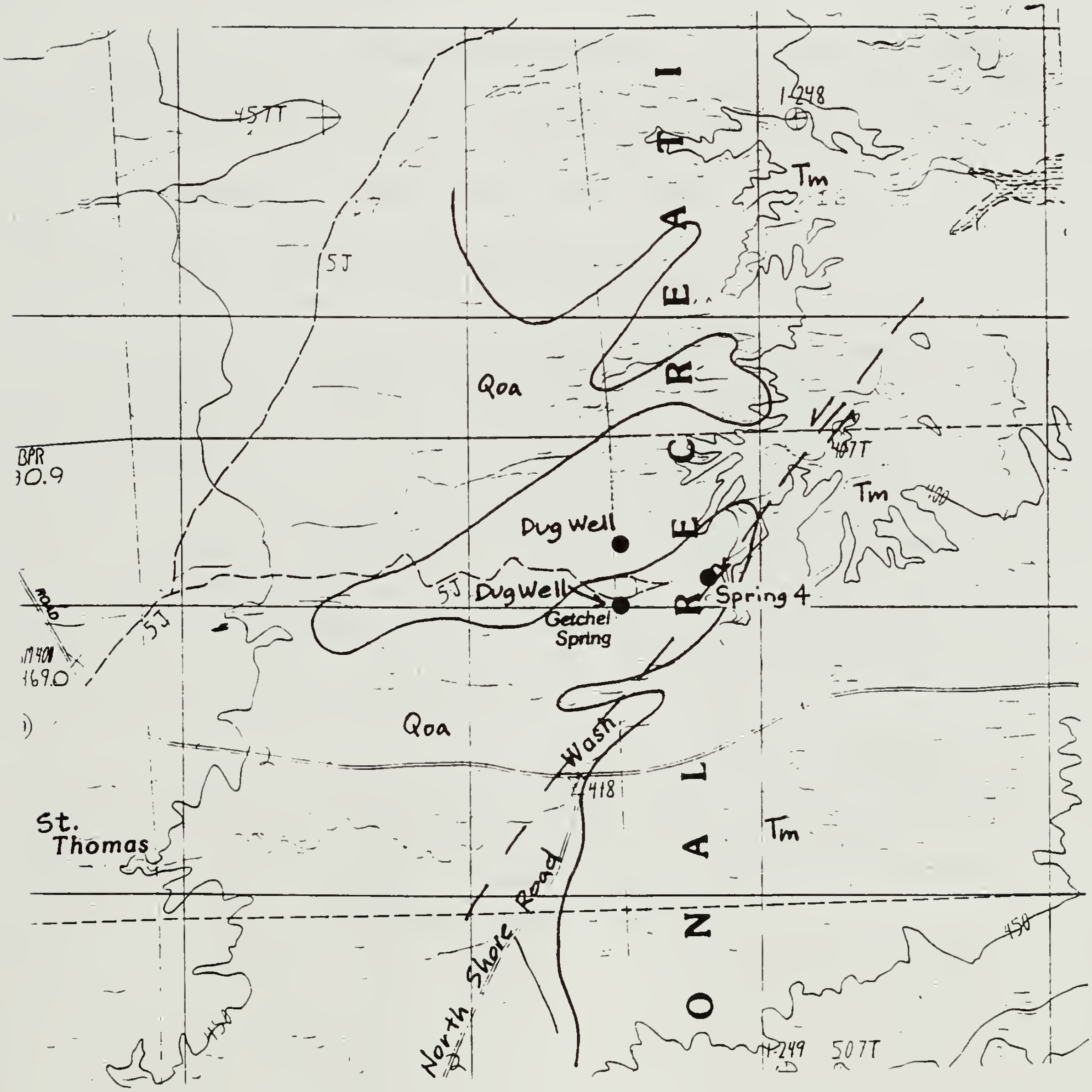
Spring 4 – Getchel Spring

Topographic base: 7.5' Valley of Fire, East Quadrangle
Geology references: Bohannon (1983)

Getchell Spring is located approximately 0.75 km northeast of the intersection of Northshore Road with the Overton Beach Road. Discharge occurs in the bottom of a 4-m-deep ravine cut into unconsolidated sands and silts of the Muddy Creek Formation. Many gypsum beds are evident within the Muddy Creek Formation near the spring. Surface flow was observed for an approximate 50 m length of the ravine on 2-9-96, although the flow was very slow to stagnant. Small amounts of vegetation were present at the orifice but very little vegetation was observed downstream.

Much of the area surrounding the spring is capped by a gyspum unit which could be in place or colluvium from above. There are no major structural features evident at ground surface, but Bohannon (1983) maps a strike-slip fault through the area, possibly related to the Lake Mead Fault System. The Rogers Spring Fault lies about 1.5 km to the southeast.

There is a dug well to the northwest of Getchel spring which contained standing water at both visits to the area (10-4-95 and 2-9-96). The well is about 2 m in diameter, 2 m deep, and filled with reeds. There is also a brick-lined cavity (cistern?) about 50 m south of the dug well and 100 m west northwest of Getchel Spring. This feature is 3 m deep and 2 m in diameter at the surface, and though it contained no water at either of our visits, it appears to be the feature labeled as Getchel Spring on the "Valley of Fire, East" 7.5' quadrangle map.


NATIONAL RECREATION AREA
Qoa
Qoa
Tm
Tm
Tm
Dug Well
Dug Well
Getchel Spring
Spring 4
Wash
North Shore Road
St. Thomas
5J
5J
5J
457T
407T
507T
1248
1249
418
450
450

Spring 5 – Unnamed uppermost spring in Valley of Fire Wash

Topographic base: 7.5' Valley of Fire, East Quadrangle
Geology references: Campagna (1990) unpublished mapping

This spring issues from several seeps at the base of the northern bank of Valley of Fire Wash, at the boundary of the recreation area. Surface flow in the wash was observed for a distance of 200 to 300 m on 3-7-96.

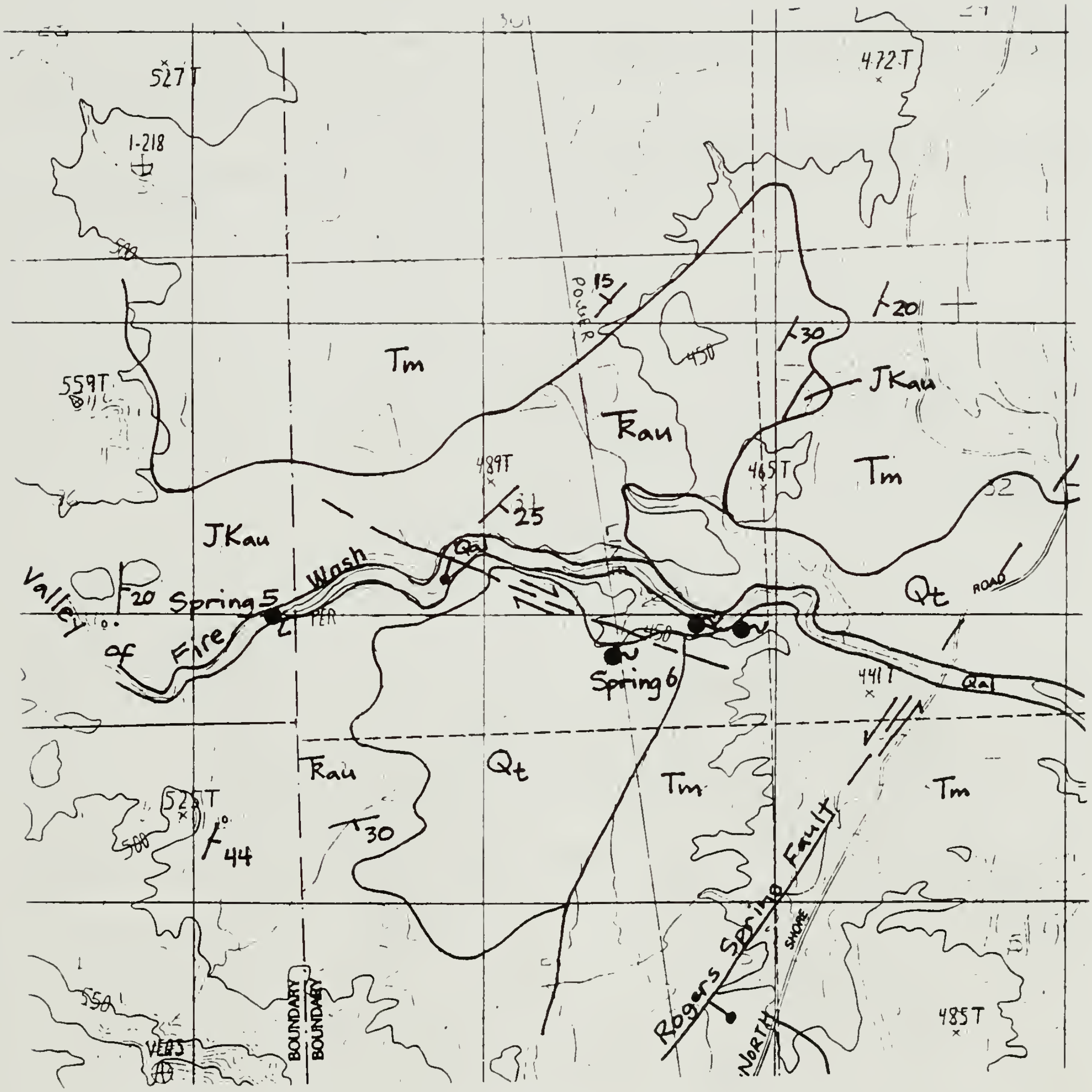
The spring is located on a fault contact between JKau on the west and TRau on the east, but is probably a result of the proximity of the contact between the Jurassic and Triassic clastic rocks with the Tertiary Muddy Creek Formation (see description of Spring 6).

Spring 6 – Unnamed upper spring in Valley of Fire Wash

Topographic base: 7.5' Valley of Fire, East Quadrangle
Geology references: Campagna (1990) unpublished mapping

Several orifices and seeps are located along the banks of the Valley of Fire Wash near the power line crossing. Surface flow from this spring area extended to within a few hundred m of North Shore Road at our 3-7-96 visit. The spring area supports a great deal of vegetation along the banks of the wash. Most of the springs and seeps are on the south side of the wash and within 5 m of the wash bottom; however, one small channel extends to the south out of the wash, originating at a spring just southwest of the power line road. Our samples were collected at this orifice, which issues from a thin veneer of Quaternary gravels on top of the Triassic Moenavi and Kayenta Formations. There appears to be considerable subsurface flow within these gravels because flow at the orifice is much lower than flow from the same channel downstream at the Valley of Fire Wash.

The spring area is located at an unconformable contact of Jurassic and Triassic clastic rocks on the west with the Tertiary Muddy Creek Formation on the east, and near the Rogers Spring Fault. The springs occur where eastward flowing groundwater meets the low-permeability barrier formed by the Muddy Creek Formation and is forced upward, possible along fault planes, to discharge points at ground surface.



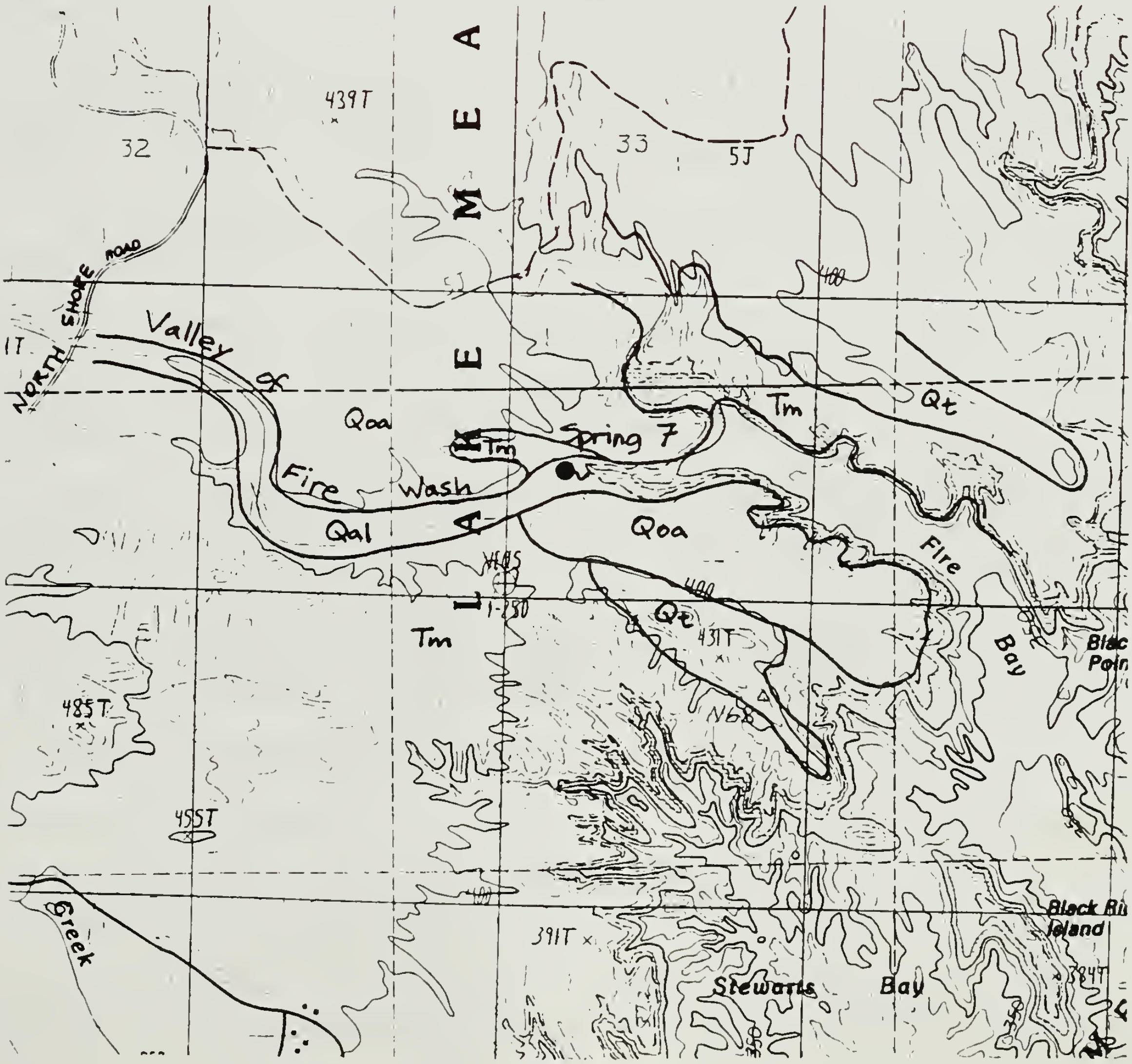
527T
1-218
472T
Tm
15
POWER
20
30
JKau
559T
Ʈ̄Rau
489T
25
465T
Tm
JKau
Qal
Valley
of
Fire
Wash
F20
Spring 5
Qt
ROAD
Spring 6
441T
Qal
Ʈ̄Rau
Qt
Tm
Tm
525T
44
30
Rogers Spring Fault
SHORE
NORTH
BOUNDARY
BOUNDARY
485T
VEGAS

Spring 7 – Unnamed lower spring in Valley of Fire Wash

Topographic base: 7.5' Valley of Fire, East Quadrangle
Geology references: Bohannon (1983)

The spring is located on the north bank of the Valley of Fire Wash and about 5 m above the base of the wash. Surface flow was evident in the wash from the spring to Lake Mead on our 2-9-96 visit, a distance of about 1 km. Seepage into the wash may be occurring along this stretch. The banks of the wash are covered by thick stands of tamarisk and other vegetation, but the main spring is in a small clearing. Several orifices and seeps are distributed along the bank. Samples were collected from the largest.

The spring issues from Quaternary Older Alluvium near an exposure of the Muddy Creek formation.



Spring 8 – Blue Point Spring

Topographic base: 7.5' Valley of Fire, East Quadrangle
Geology references: Bohannon (1983), Campagna and Aydin (1994)

Blue Point Spring is at the base of the Muddy Mountains, 350 m west of North Shore Road. The spring issues from colluvium about 10 m horizontally from the nearest limestone exposure, and into a 3-m-deep ravine. The surface flow forms Slim Creek, which flows southeast toward Stewarts Point and Lake Mead. Parts of Slim Creek flow underground in locations where the gypsum–rich soils have been dissolved. The spring orifice is surrounded by thick acacia and other vegetation. Samples were collected at the orifice.

The spring is located at the point of intersection of the Rogers Spring Fault and the older west-northwest-trending Arrowhead Fault.

Spring 9 – Unnamed spring 0.8 km south of Spring 8

Topographic base: 7.5' Valley of Fire, East Quadrangle
Geology references: Bohannon (1983), Campagna and Aydin (1994)

Spring is located approximately 50 m east of a culvert under North Shore Road. No surface flow was evident although saturated soils support a dense stand of cat tails and other vegetation, including several cottonwood trees, in an area about 20 m wide.

Spring issues from unconsolidated and partially consolidated red and tan silts, with interbedded sand, pebbles, and gypsum.

Spring 10 – Unnamed spring 0.8 km southeast of spring 9

Topographic base: 7.5' Valley of Fire, East Quadrangle
Geology references: Bohannon (1983), Campagna and Aydin (1994)

Spring is located about 0.8 km southeast and in the same wash channel as Spring 9. Discharge is diffuse and widely-distributed across the base of the wash channel (25 to 30 m wide), although several small (less than 1 m across and 0.1 m deep) channels have been developed. Dense vegetation throughout seep area, including mesquite, tamarisk, and reeds. Our discharge measurement was made upstream of the most diffuse flow and therefore does not account for the diffuse discharge, which is the majority of the discharge from this spring.

Spring issues from Quaternary terrace deposits.

Spring 11 – Rogers Spring

Topographic base: 7.5' Valley of Fire, East Quadrangle
Geology references: Bohannon (1983), Campagna and Aydin (1994)

Rogers Spring is 300 m west of the North Shore Road at the base of the Muddy Mountains. The spring issues from brecciated limestone into a manmade pool having a

diameter of about 25 m. The orifice is below the surface of the pool. Overflow from the pool enters Rogers Wash and flows southeast across basin-fill deposits about 3 km to where it enters Lake Mead. Rogers Spring is the largest spring in the study area, with a relatively constant discharge of 2,550 L/min measured since 1985 (USGS, 1996). Samples were collected by submerging and opening the sample bottles below the pool surface at the spring orifice.

The spring is located on the Rogers Spring Fault, a major strike-slip fault in the Lake Mead area. The fault separates lower Paleozoic carbonate rocks of the Muddy Mountains on the west from Quaternary and Tertiary basin-fill deposits to the east. The low permeability basin fill is a barrier to groundwater flow that causes the Rogers Spring Fault to act as a conduit for flow from depth within the carbonates. Four springs issue directly from the fault and several more issue from the basin fill between the fault and Lake Mead.

Rogers Spring is at a step-over in the main Rogers Spring Fault. Fracture density increases near step-over zones in extensional terrains, increasing the potential for groundwater flow paths.

Spring 12 – Scirpus Spring

Topographic base: 7.5' Echo Bay Quadrangle
Geology references: Bohannon (1983), Campagna and Aydin (1994)

Scirpus Spring is 550 m southwest of Rogers Spring. The spring consists of a primary pool 3 m long and 0.5 m wide that is surrounded by very thick reeds, shrubs, and grape vines. No surface flow was evident when this spring was visited (2-7-96). However, abundunt phreatophytes grow in the ravine below the spring indicating evapotranspiration is a major component of spring discharge. Samples were collected from the pool.

The spring is located along the Rogers Spring fault and issues from brecciated limestone about 25 m downslope from bedded limestone of the Muddy Mountain front.

Spring 13 – Corral Spring

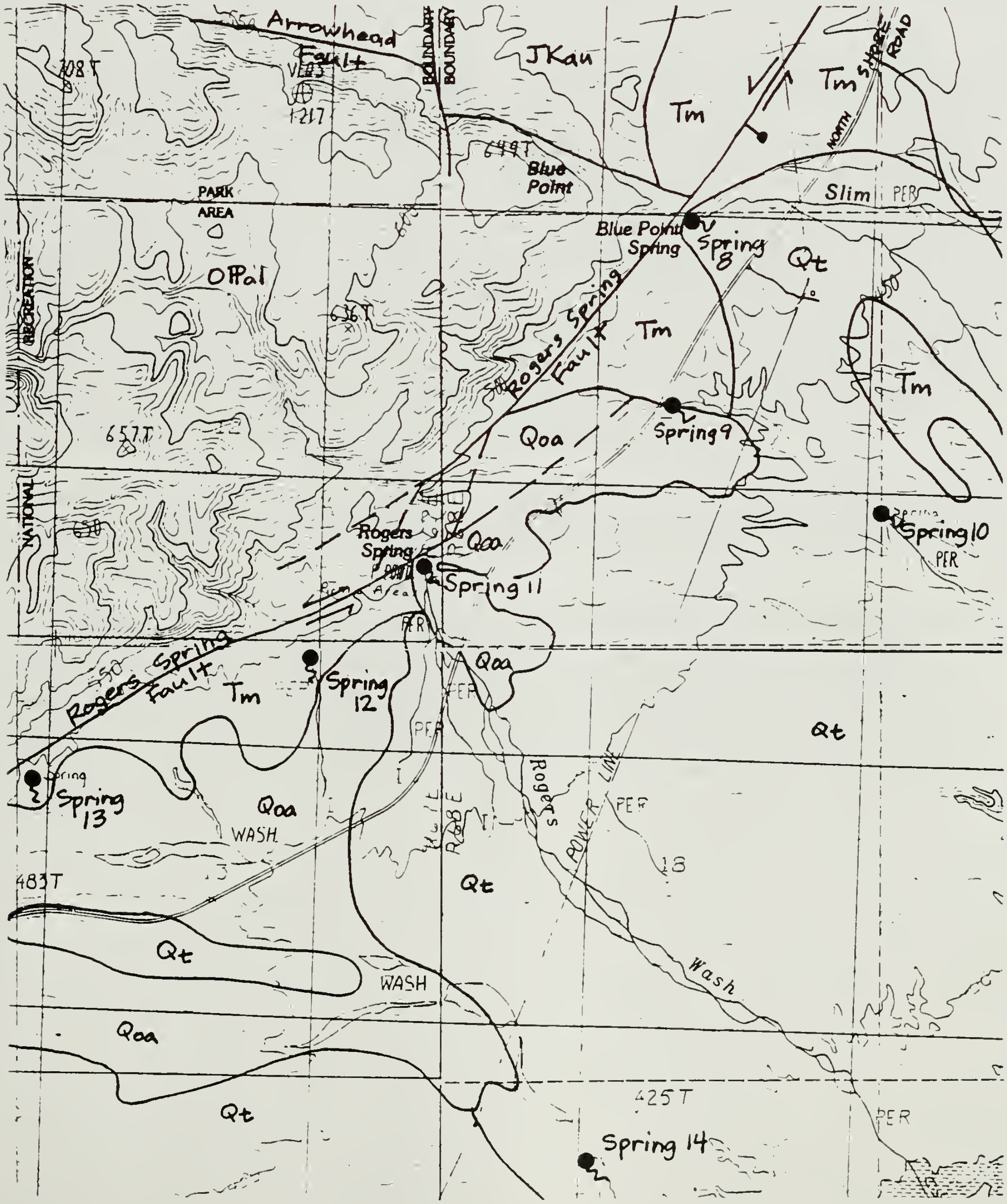
Topographic base: 7.5' Echo Bay Quadrangle
Geology references: Bohannon (1983), Campagna and Aydin (1994)

Corral Spring is the southernmost spring on the Rogers Spring Fault and is located about 1.7 km southwest of Rogers Spring. The spring issues from colluvium in a steep canyon that extends into the limestone of the Muddy Mountain front. The spring consists of several isolated seeps and small pools distributed along a 100 m length of the base of the canyon. Little surface flow was evident, however. This area supports a great deal of vegetation, suggesting that evapotranspiration is a major component of spring discharge. Samples were collected from the highest pool, which was about 4 m long and 2 m wide, and half filled with reeds, at our 2-7-96 visit.

Spring 14 – Unnamed spring northwest of Rogers Bay

Topographic base: 7.5’ Echo Bay Quadrangle
Geology references: Bohannon (1983)

Spring issues from wash bottom as seeps. Discharge measurement made approximately 20 m downstream from highest seep. Grasses and mesquite surround the spring area, but there is considerably less vegetation than at other springs in the North Shore Spring Complex. Spring issues from Quaternary terrace deposits.



Spring 15 – Bitter Spring

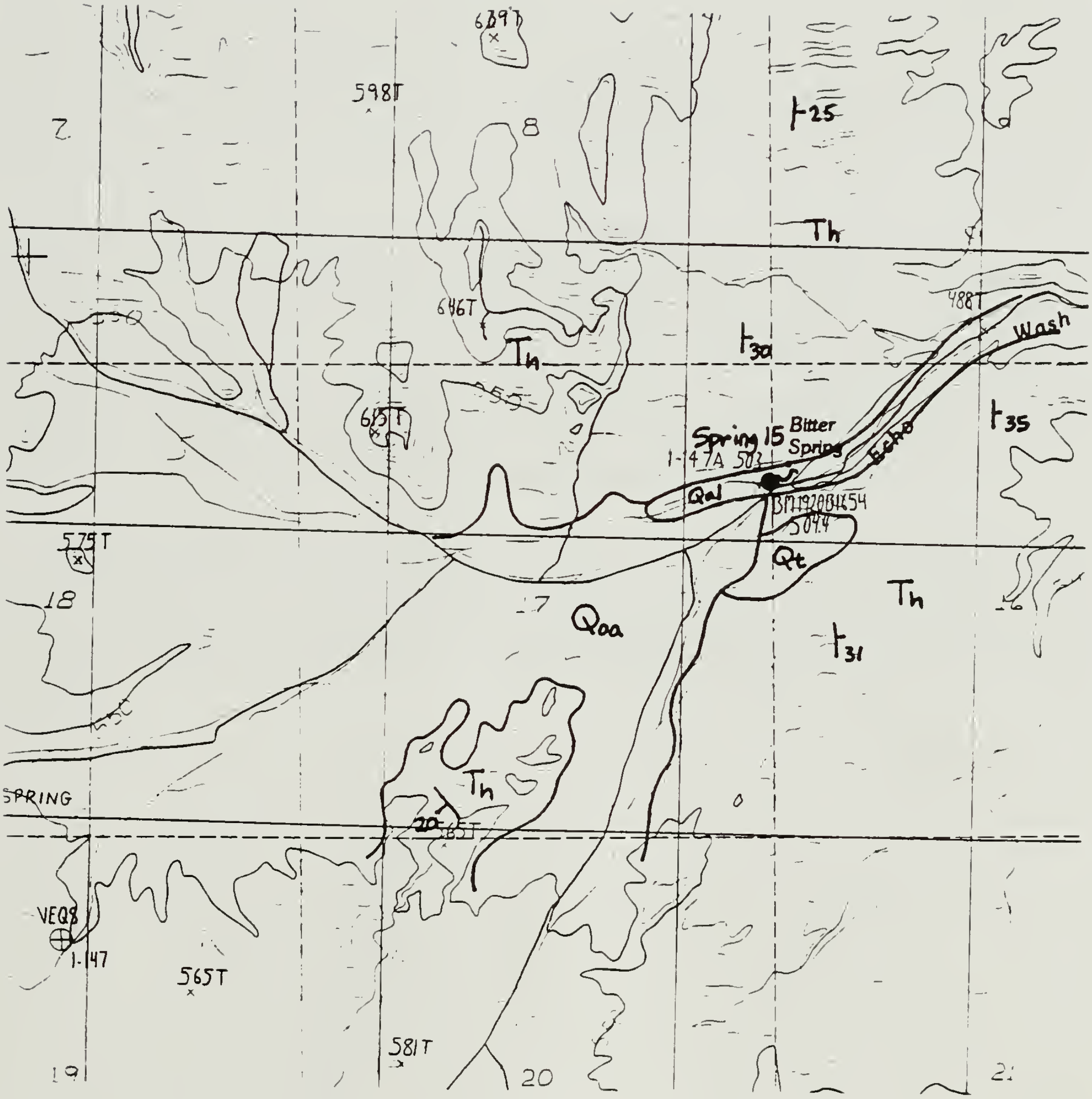
Topographic base: 7.4' Bitter Spring Quadrangle
Geology references: Bohannon (1983)

Bitter Spring is located in Echo Wash at the eastern margin of Bitter Spring Valley. The spring consists of relatively diffuse flow issuing from coarse sand and gravel alluvium in the center of the wash, approximately 0.5 km east and downstream of the channel knick point, which is composed of consolidated Older Alluvium. At the spring, the wash channel is incised in clastic and associated chemical and tuffaceous rocks of the Thumb member of the Tertiary Horse Springs Formation, which dips 25 to 35 degrees east.

Surface drainage to Bitter Spring originates in Bitter Spring Valley directly to the west, and White Basin to the northwest of that. Bitter Spring Valley is composed of approximately 1,500 m of Horse Spring Formation and is covered by Pleistocene alluvium, Pleistocene terrace deposits, and Thumb Member. Bohannon (1983) hypothesizes a section of Paleozoic carbonate rocks below the Thumb. The Bitter Spring Valley margins are composed of Horse Spring Formation to the north and west (Bitter Ridge), and autochthonous Triassic and Permian formations to the south (Razorback Ridge and Pinto Ridge). The subsurface geology of White Basin is similar, but the surface geology differs in that Thumb Member is not exposed and large deposits of Miocene Red Sandstone are present. On the west, White Basin is bordered by Autochthonous Jurassic, Cretaceous, and Triassic rocks; and on the north by Allochthinous lower Paleozoic rocks (Muddy Mountains).

Bitter Spring is located near the eastern terminus of the Borax Fault, and the southern end of East Longwell Ridge; however, the spring does not appear to be directly related to any major structural feature.

Surface flow from the spring is evident, but discontinuous, over a 300 m distance below the orifice, and is accompanied by dense stands of phreatophytes (primarily tamarisk). It is likely that our measurement of discharge at Bitter Spring represents only a small portion of the total spring flow when compared to underflow in the wash sediments, evaporation from the surface channels, and transpiration from plants. Samples were collected from the highest discharge point.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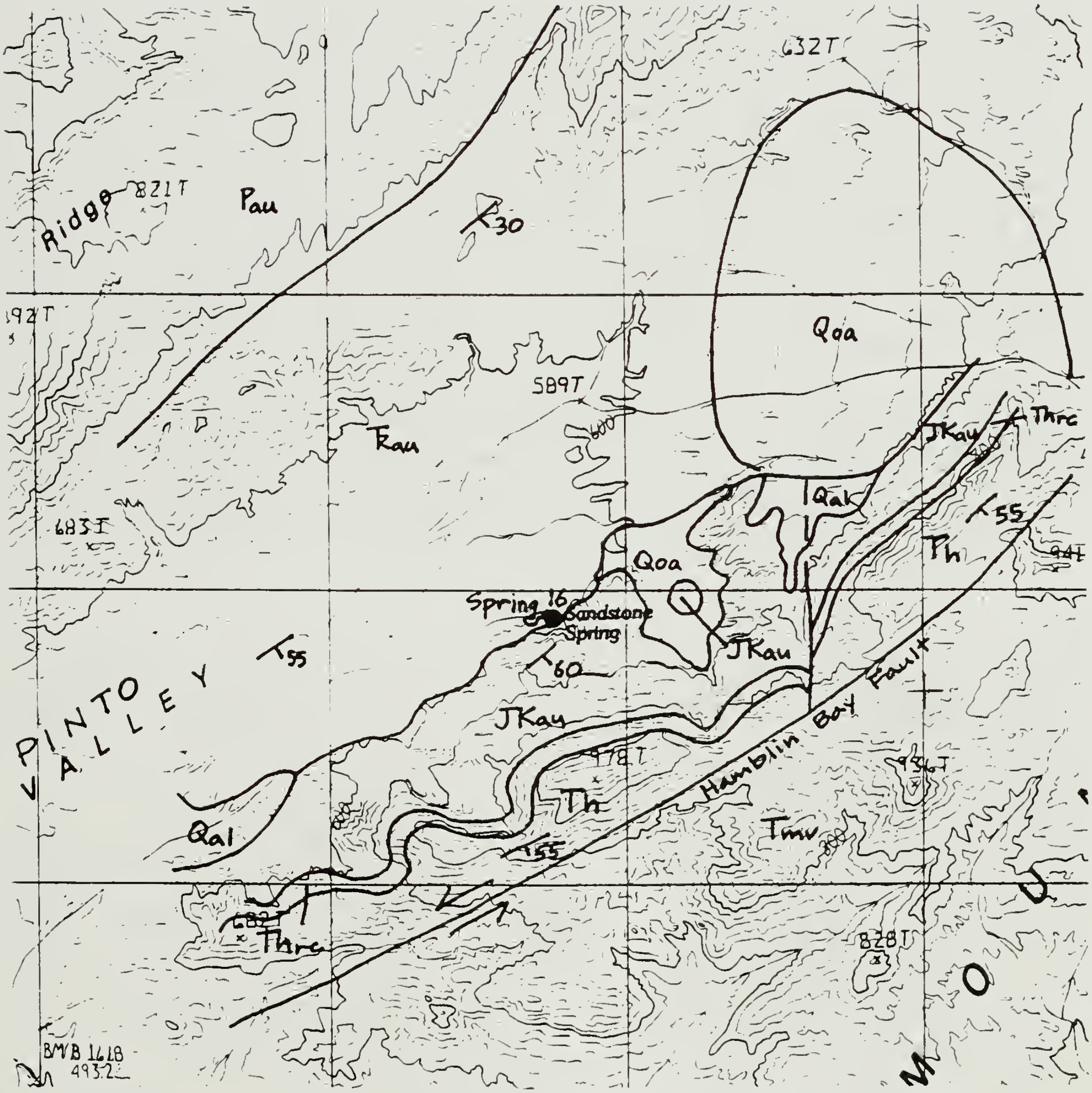
Spring 16 – Sandstone Spring

Topographic base: 7.5' Boulder Canyon Quadrangle
Geology references: Bohannon (1983)

Sandstone Spring is located at the southeast margin of Pinto Valley, and northwest of the Black Mountains. The spring issues at the base of a cliff composed of Aztec Sandstone, which is several hundred meters in height, and into a single pool having a diameter of approximately 2 m. A steel pipe leads from the pool to a steel tank about 20 m downhill from the spring, but the tank contained no water at either of our visits (10-3-95 and 2-7-96). Samples were collected from seepage into the pool. Longwell noted the existence of this spring in his (date?) report and described its quality and quantity as sufficient for watering horses.

Large surface runoff events are evident through the spring area as indicated by the wash channel that cuts into the alluvial fan deposits northwest of the spring and then extends downstream from the spring, and the eroded surface of the sandstone on the cliff face above the spring. Surface flow of this type may serve to recharge shallow sediments and provide temporary "spring discharge" during wet periods; however, atmospheric tritium was not detected in a sample collected 2-7-96 indicating that flow paths are long and that recent recharge was not a major component of spring discharge at that time.

Sandstone Spring is located at a contact of the Jurassic Aztec Sandstone with the underlying Triassic Moenave and Kayente Formations (clastic, nearshore marine and nonmarine rocks). The contact trends N 60 E and dips 60 degrees to the southeast. Discharge at Sandstone Spring may be related to nearly vertical fractures in the Aztec that trend north-south.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Spring 16
Sandstone Spring
JKau
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PINTO VALLEY
JKay
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Hamblin Bay Fault
936T
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682T
Thrc
828T
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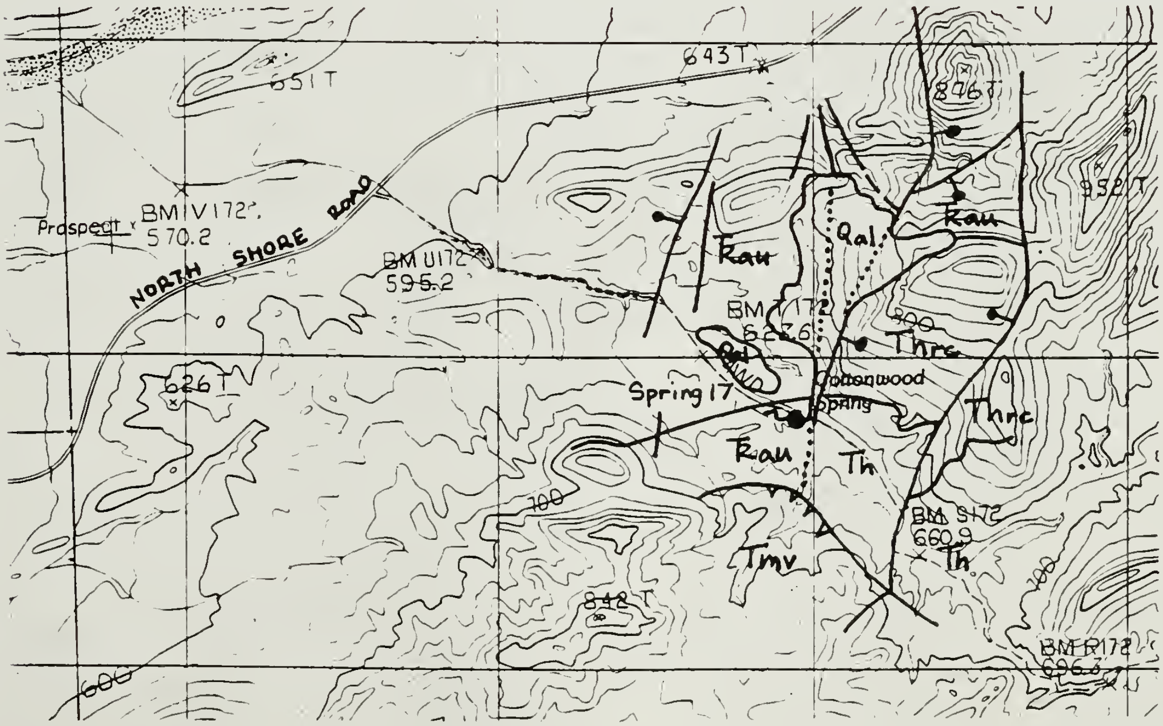
MOU

Spring 17 – Cottonwood Spring

Topographic base: 7.5' Callville Bay Quadrangle
Geology references: Anderson (1973), Campagna (1990) unpublished mapping

Cottonwood Spring is located approximately 2.5 km north of Hamblin Mountain and 1.5 km southeast of North Shore Road, in a wash channel that is tributary to Callville Wash. It appears that spring discharge has in the past occurred from alluvial sediments in the northwest-trending wash channel just downstream of a 3-m high dry waterfall. There are two cottonwood trees located here and evidence of several holes dug by bighorn sheep, burros, or horses in search of water. However, surface discharge was not evident at this location during either of our visits (10-3-95 and 2-6-96). The only discharge evident from the area was from a steel pipe into a metal tank about 40 m southwest of the cottonwood trees. On 10-3-95, the tank was only partially full, indicating some leakage through the sides and insufficient spring discharge to keep it completely full. On 2-6-96, the tank was completely full and overflowing, suggesting that discharge was somewhat greater than observed during the 10-3-95 visit. The tank is useful to wildlife, as we observed several desert bighorn sheep during the 10-3-95 visit. Samples were collected from the pipe as it discharged into the tank.

The orifice is located at a north-south-trending fault contact of the Tertiary Rainbow Gardens basal conglomerate (on the east) with the Triassic upper red unit of the Moenkopi Formation (on the west). The basal conglomerate is approximately 10 to 20 m thick in the area of the spring. The alluvium filling the wash is probably less than 10 m thick.

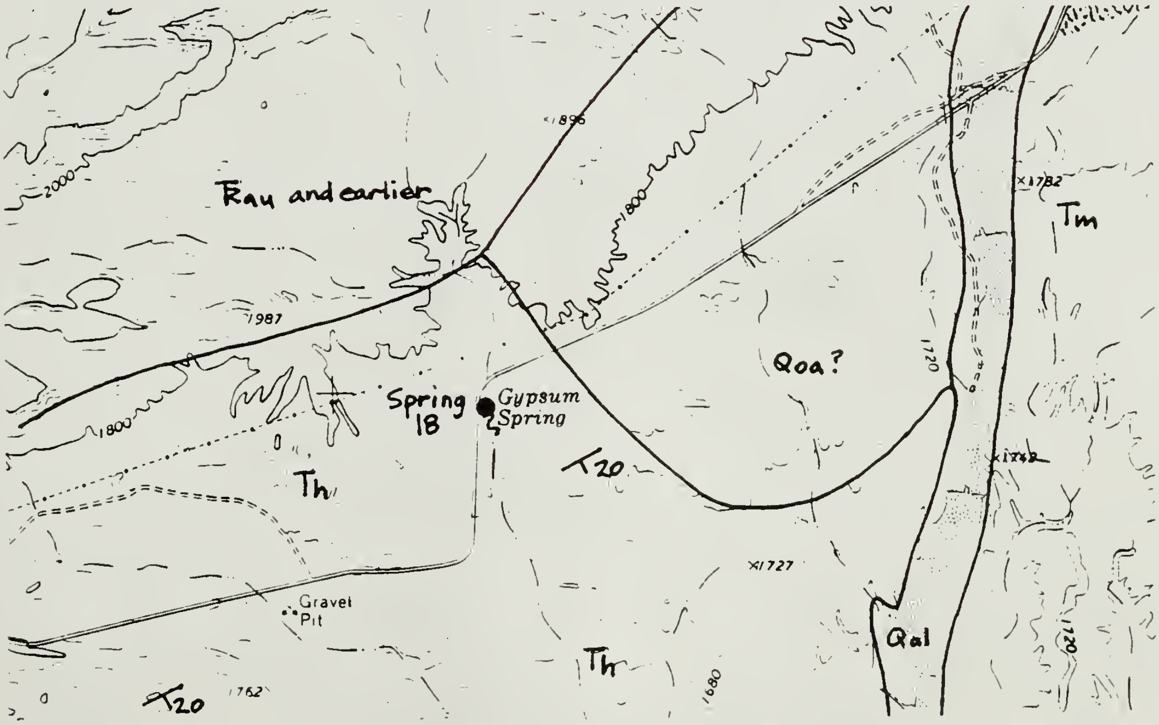


Spring 18 – Gypsum Spring

Topographic base: 7.5' Frenchman Mtn. Quadrangle
Geology references: Bohannon (1978), Longwell *et al.* (1965)

The spring is located approximately 6 km east southeast of Sunrise Mountain and about 1.5 km southwest of Gypsum Cave. The surface discharge is characterized by several seeps and pools in a 3-m-deep, north-south-trending wash channel. The pools were less than 1 m in diameter at both visits (10-2-95 and 2-6-96) and surface flow was present for less than 15 m downstream of the highest orifice. Very dense stands of tamarisk and reeds surround the orifice and line the banks of the wash channel. Samples were collected from surface flow as it emerges from dense vegetation near the orifice.

Gypsum Spring issues from gypsum beds of the Thumb Member about 0.5 km south of a ridge composed of Triassic and older rocks. The spring discharge appears to be controlled by the intersection of the water-bearing unit with land surface; no structural control is evident. Although the elevation of the spring is lower than water levels in the carbonate aquifer to the north in Dry Lake Valley, stable isotopic data indicate that the carbonates are not the source for discharge at Gypsum Spring. Rather, this spring plots in the region of low-elevation precipitation which indicates that its flow was recharged locally. As with other locally-derived springs, the absence of detectable atmospheric tritium in the spring water indicates that despite the local origin, travel times are long and the discharge does not simply represent discharge of groundwater recharged during recent precipitation events.

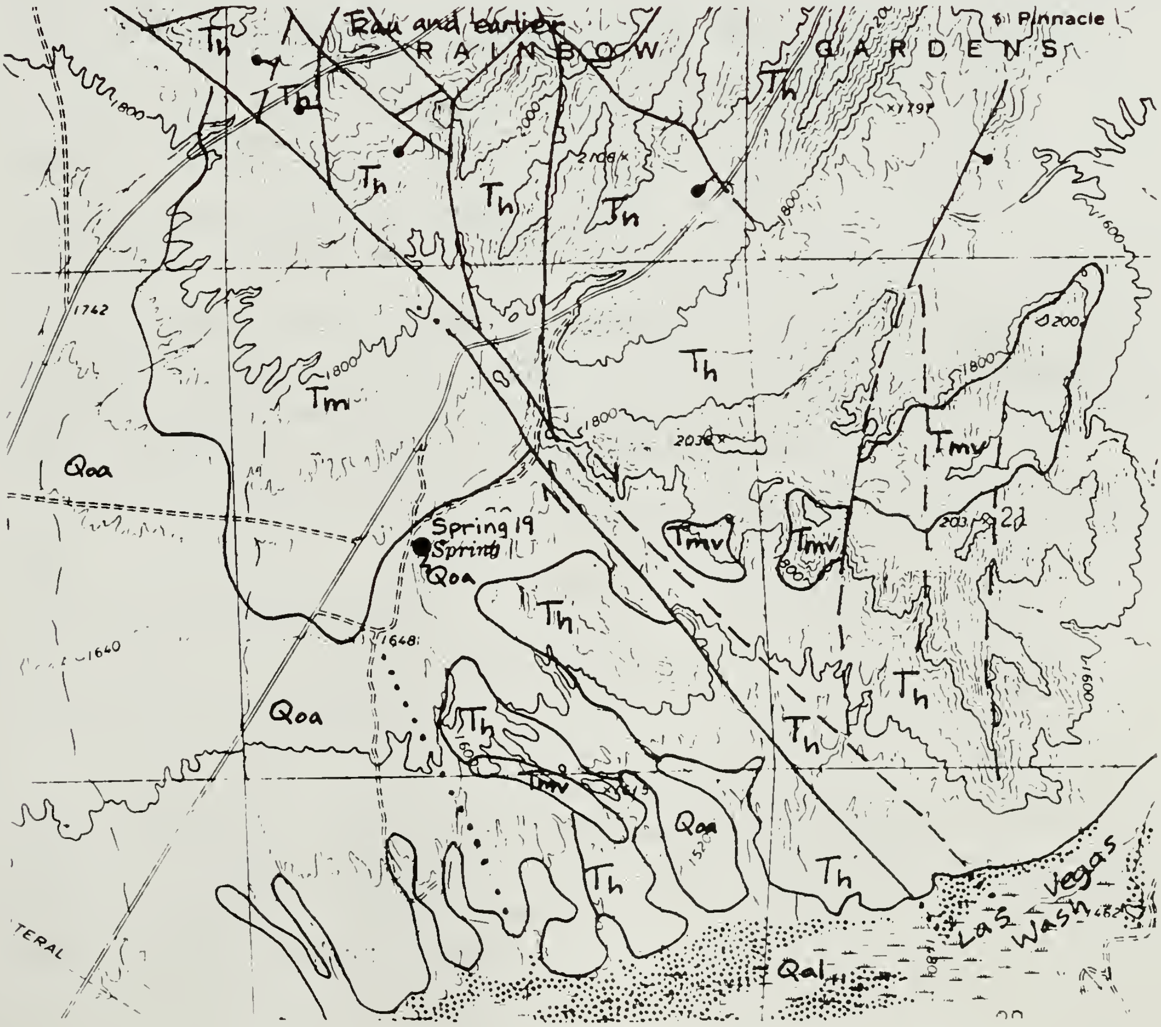


Spring 19 – Unnamed spring south of Rainbow Gardens

Topographic base: 7.5' Henderson Quadrangle
Geology references: Bell and Smith (1980)

The spring is located at the southern end of Rainbow Gardens and about 1.75 km north of Las Vegas Wash. The spring issues from Quaternary alluvial fan deposits in the bottom of a 5-m-deep wash channel incised in the fan surface. The surface flow originates from a single orifice but the area around the spring supports thick tamarisk, mesquite, and grasses; presumably related to shallow groundwater throughout the area. Flow is at the surface for approximately 10 m before it infiltrates into the alluvial deposits. A pipe and circular concrete tank suggest that the spring has been utilized as a water supply in the past, but both are presently filled with sediment.

The spring is near a step-over in a major northwest-trending strike-slip fault (Bell and Smith, 1980). The fault forms a boundary between the Tertiary Horse Spring Formation to the northeast and Quaternary alluvial fan deposits and Tertiary Muddy Creek Formation to the southwest.



Spring 20 – Pupfish Spring

Topographic base: 7.5' Hoover Dam Quadrangle
Geology references: Mills (1994)

The main spring is 30 m upslope of a concrete tank, which is located on the west side of the Lower Portal Road, just above the tunnel to the base of Hoover Dam. The pool issues as a 6-m-high waterfall into a 4-m-diameter pool. The top of the waterfall was inaccessible, so samples were collected from the pool. Dense vegetation surrounds the pool and the channel that leads to the river. Measurements of flow rate were made just above where the channel enters the river. In addition to the main spring, there are numerous seeps along the cliff face between the spring and the river.

Spring 21 – Arizona Hot Spot

Topographic base: 7.5' Hoover Dam Quadrangle
Geology references: Mills (1994)

Several seeps and springs issue from the Arizona side of the river, about 1.6 km downstream of Hoover Dam. The largest of these is the furthest downstream and is located almost directly across the river from the mouth of Goldstrike Canyon. Samples were collected from an orifice at the margin of a talus slope, about 10 m above the river.

The springs issue from Miocene Patsy Mine volcanics (undifferentiated).

Spring 22 – Sauna Cave

Topographic base: 7.5' Hoover Dam Quadrangle
Geology references: Mills (1994)

Sauna Cave is a shaft mined into the wall of Black Canyon on the Nevada side of the river, and is located 1.4 km below the dam. Groundwater discharges at the back end of the shaft and flows out of the mouth. Samples were collected at the point of discharge at the back end of the shaft. Flow measurements were made at the mouth.

The shaft is mined into the Boulder City Pluton and intersects a north-south-trending fault.

Spring 23 – Nevada Hot Spring

Topographic base: 7.5' Hoover Dam Quadrangle
Geology references: Mills (1994)

Several springs issue from the floor and walls of Goldstrike Canyon about 600 m upstream from the river. Although most of the discharge into the channel is relatively diffuse, we sampled from a point orifice at the base of the north wall, about 100 to 150 m below the highest point of discharge. Discharge measurements were conducted about 75 m upstream from the concrete dam at the riverbank.

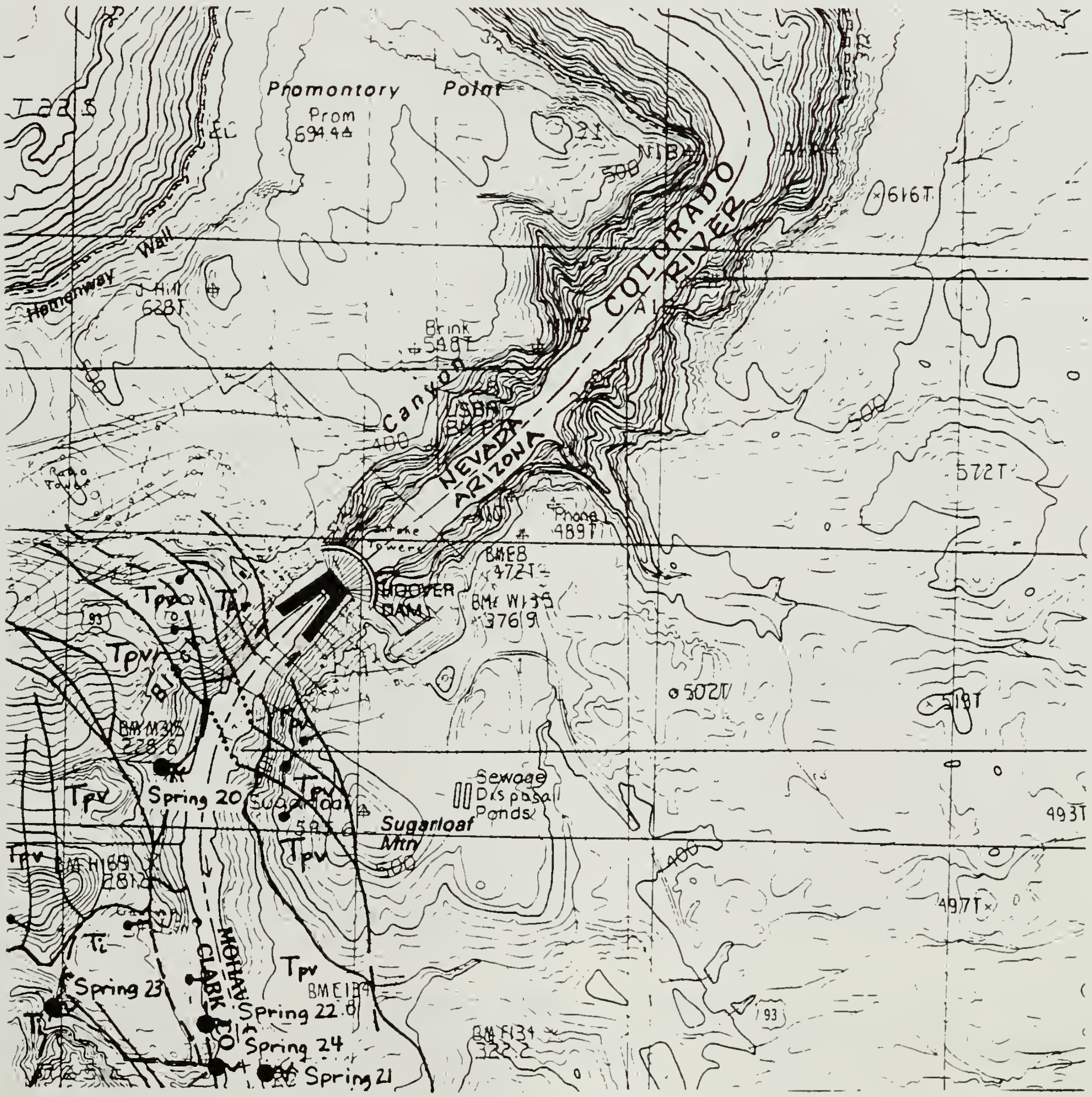
The spring issues from a north-south-trending high angle fault in the Miocene Boulder City Pluton.

Spring 24 – Nevada Hot Spot

Topographic base: 7.5' Hoover Dam Quadrangle
Geology references: Mills (1994)

The spring issues into a small cove on the Nevada side of the river, about 1.6 km below Hoover Dam. There are two main orifices above the river, but many seeps and drips, and possible subsurface discharge to the river. Large ferns overhang the river.

The spring issues from a north-south-trending high angle fault in the Miocene Boulder City Pluton.



Spring 25 – Palm Tree, Hot

Topographic base: 7.5' Ringbolt Rapids Quadrangle
Geology references: Anderson (1978)

This spring is located about 100 m from the river in a ravine that meets the river about 2.25 km below the dam. The spring issues as diffuse flow from the banks of the ravine. A cold spring (Palm Tree Cold) issues about 100 m upstream of the hot spring. The floor of the ravine is covered by very dense tamarisk. The combined surface flow of the warm and cold springs extends down the ravine to the river.

The spring issues from Miocene Patsy Mine volcanics (undifferentiated) near a northwest trending right lateral strike-slip fault.

Spring 26 – Palm Tree, Cold

Topographic base: 7.5' Ringbolt Rapids Quadrangle
Geology references: Anderson (1978)

This spring is located about 200 m from the river in a ravine that meets the river about 2.25 km below the dam. A warm spring (Palm Tree Hot) issues about 100 m below the cold spring. The floor of the ravine is covered by very dense tamarisk, making access to the cold spring very difficult. An area of reeds grows just above the highest orifice of the cold spring, where the ravine widens and the floor flattens. Surface flow extends down the ravine to the warm spring, and the combined flow extends to the river.

The spring issues from Miocene Patsy Mine volcanics (undifferentiated) near a northwest trending right lateral strike-slip fault.

Springs 28 and 29 – Boy Scout Canyon

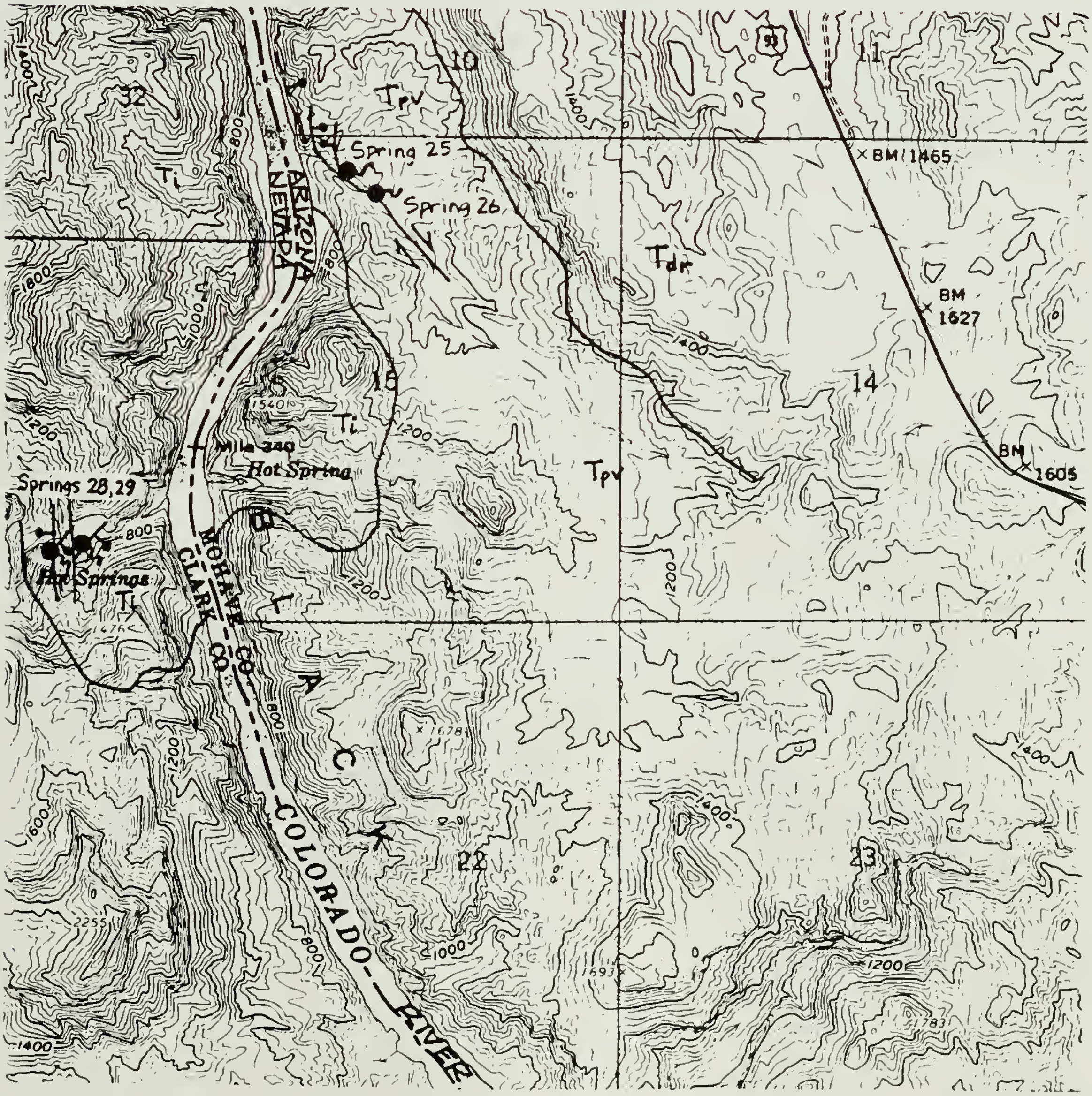
Topographic base: 7.5' Ringbolt Rapids Quadrangle
Geology references: Anderson (1978)

There are a number of springs and seeps in Boy Scout Canyon, and a wide variety of temperatures. Boy Scout Canyon is on the Nevada side and meets the river about 3.5 km below the dam. The lowest point of discharge is about 400 m up the canyon from the river. At this location, cold water discharge forms a waterfall about 12 m high and warm water discharge issues from seeps just above the floor of the canyon. The highest area of warm discharge occurs as seepage from an overhanging wall about 50 m upstream from the springs just described. The surface flow above this point is cold and passes over several waterfalls. Samples were collected of both the warm and cold discharge. Note that despite their difference in temperature, these springs have very similar geochemical and isotopic composition.

Although the discharge rate from this spring is relatively high, only a small fraction of the surface flow reached the river on our visit of 2-2-97. Several reaches of the channel

carried no surface flow. The discharge measurement was made at the farthest downstream location of channel flow over a bedrock bench.

The springs issue from the Miocene Boulder City pluton at points where near vertical, north-south-trending faults intersect from below an unconformable barrier. This unconformity appears to act as a "ceiling", preventing further flow within the plutonic rocks.

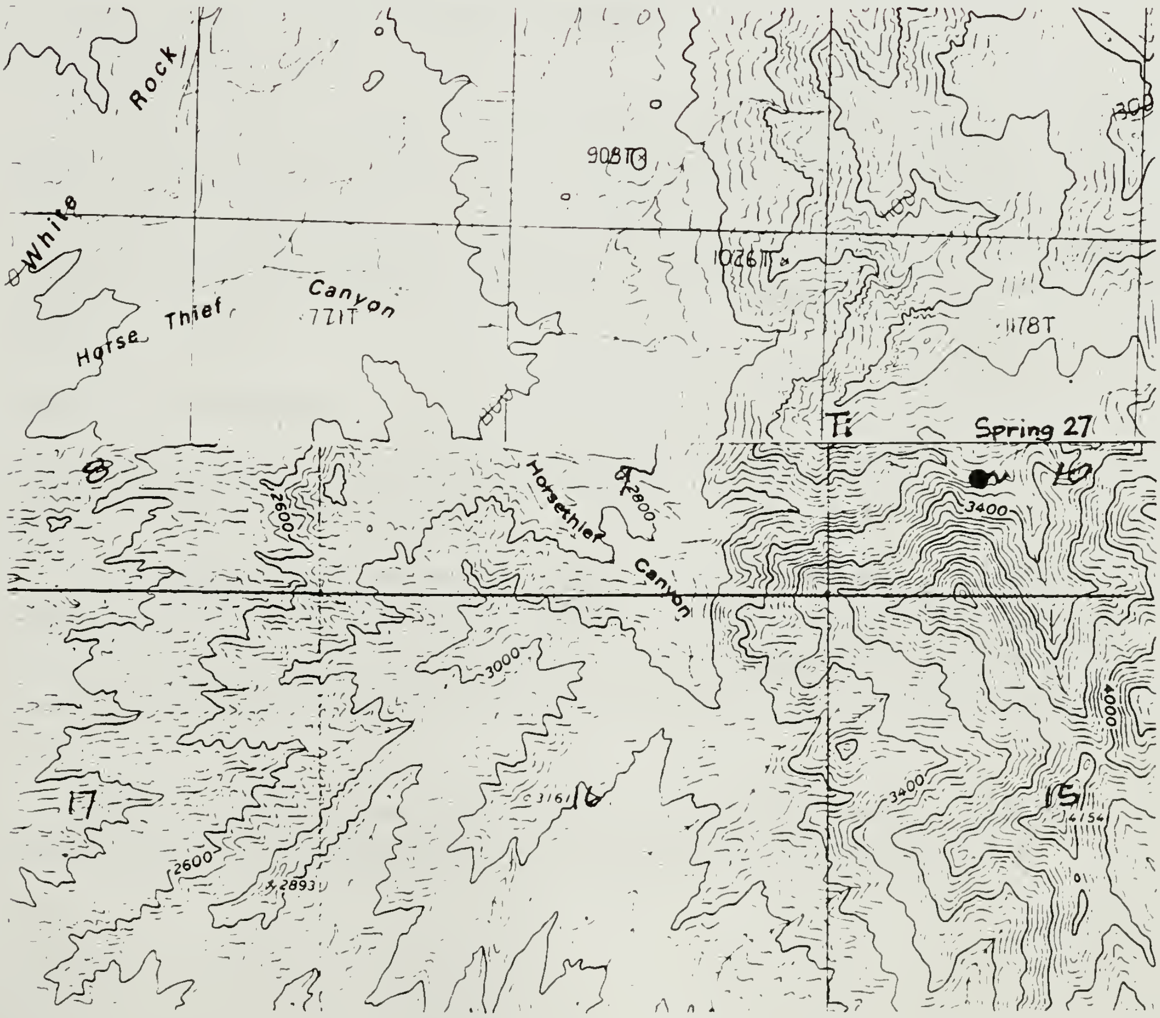


Spring 27 – Unnamed Spring in Horsethief Canyon

Topographic base: 7.5' Ringbolt Rapids Quadrangle
Geology references: Anderson (1978)

Horsethief Canyon extends into the west side of Mount Wilson of the Black Mountains in Arizona. Several springs and seeps occur in the canyon above a dry waterfall, supporting a wide variety of vegetation. At the time of our visit (2-3-97), the highest flow rate occurred about 1 km upstream from the waterfall, and supported a stand of cottonwood trees, reeds, and other vegetation. Surface flow was discontinuous over a total length of several hundred meters. Flow was on the surface in reaches where bedrock benches formed the base of the canyon, or where the alluvial deposits were thin. In other reaches, flow presumably occurs within the alluvial deposits.

The spring issues from Tertiary intrusive granite of the Wilson Ridge pluton (described by Anderson *et al.*, 1972).



Spring 30 – Arizona Hot Spring

Topographic base: 7.5' Ringbolt Rapids Quadrangle
Geology references: Anderson (1978)

Arizona Hot Spring is located in a dramatic slot canyon that meets the river just downstream of Ringbolt Rapids, and about 6.6 km downstream of the dam. The spring issues into several manmade pools that are located about 300 m up the canyon from the river. The canyon walls near the pools are nearly vertical and 2 to 3 m apart at the base. Above the pools, the canyon opens up and the walls slope gently away from the alluvium-filled channel. Surface flow extends about 150 m down the canyon from the pools, much of it in a bedrock channel, but infiltrates when the channel passes over alluvial gravels.

The spring issues from Miocene Patsy Mine volcanics (undifferentiated) near a northwest trending right lateral strike-slip fault. This fault is offset by a north-south-trending normal fault and the spring issues from near the intersection of the two faults.

Spring 31 – Unnamed cold spring near Arizona Hot Spring

Topographic base: 7.5' Ringbolt Rapids Quadrangle
Geology references: Anderson (1978)

The spring is located about 20 m up the canyon from the highest (man-made) pool of Arizona Hot Spring. Above this spring, the canyon is wide, the walls slope gently, and the floor is covered by alluvium. Below the spring, the canyon narrows dramatically (forming a "slot canyon"), the walls are nearly vertical, and the floor is scoured bedrock. The flow issues from alluvium in the base of the canyon, just above the point where the channel enters the slot canyon.

Spring 32 – Nevada Falls

Topographic base: 7.5' Ringbolt Rapids Quadrangle
Geology references: Anderson (1978)

Nevada Falls spring is located in a small cove on the Nevada side, approximately 8.2 km below the dam. Surface flow originates about 11 m above the gravel bank of the river, and drops to the river in a series of waterfalls. Only the highest pool contains vegetation. Samples were collected from the second pool up from the riverbank, which is about 3 m above the bank.

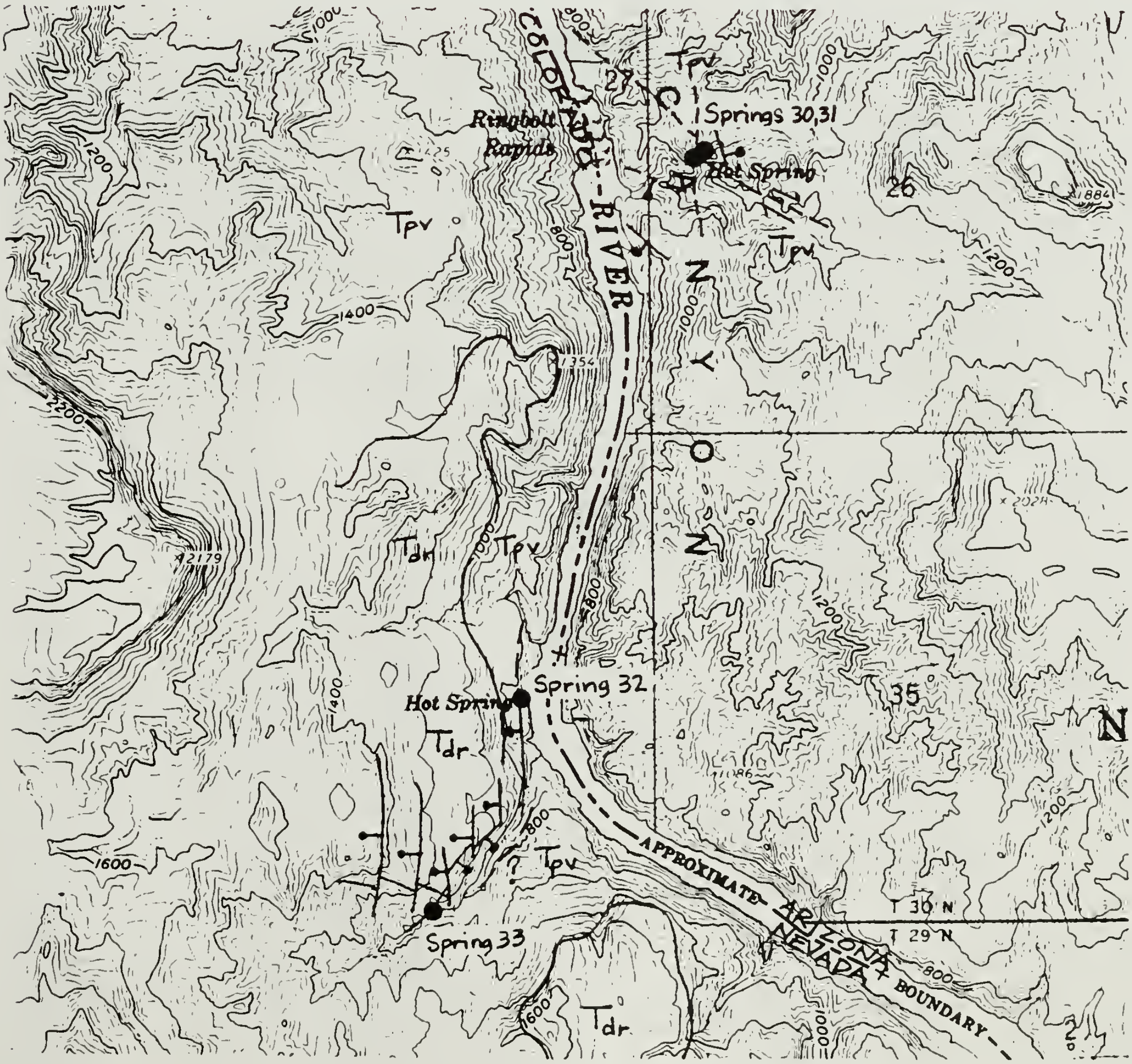
The flow issues from a north-south trending fault in the Miocene Patsy Mine volcanics (undifferentiated), about 100 m east of a contact with Tertiary Mount Davis lavas.

Spring 33 – Bighorn Sheep Spring

Topographic base: 7.5' Ringbolt Rapids Quadrangle
Geology references: Anderson (1978)

Bighorn Sheep spring is located in a steep-sided canyon that meets the river 8.4 km below the dam. The main orifice forms a 5-m-high waterfall on the north side of the canyon, about 600 m up the canyon form the river. Because the orifice was inaccessible, the samples were collected near the base of this waterfall. Additional discharge occurs at several small seeps located upstream of the main orifice, all discharging from the north wall of the canyon. Surface flow is present in the channel to within 100 m of the river, but did not reach the river on our 2-2-97 visit. Dense stands of tamarisk extend from the orifice all the way to the river.

The spring issues from a northeast-trending fault in the Tertiary lavas (Mount Davis Volcanics).

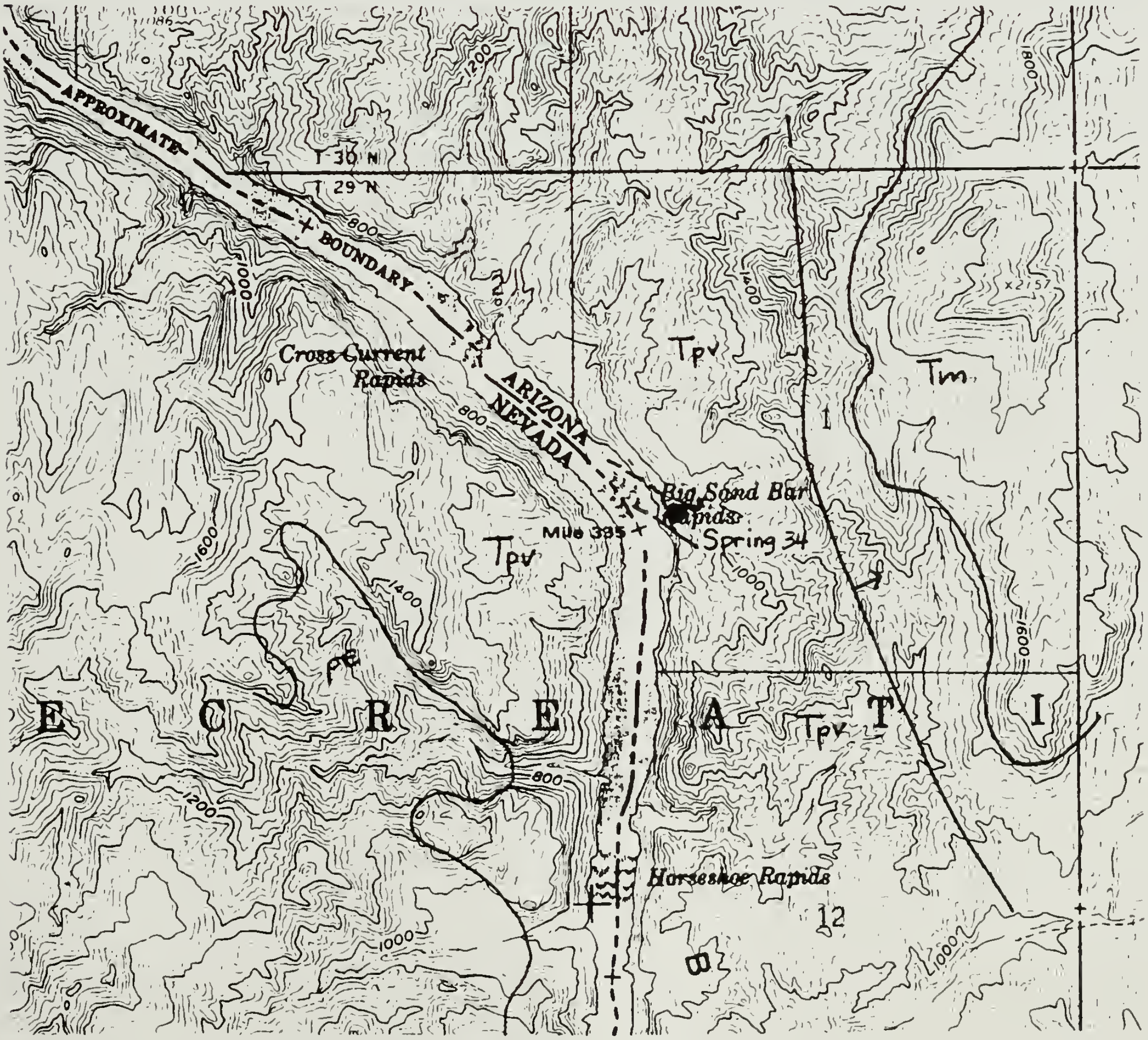


Spring 34 – Arizona Seep

Topographic base: 7.5' Ringbolt Rapids Quadrangle
Geology references: Anderson (1978)

Arizona seep is 11.3 km below Hoover Dam, on the Arizona side of the river. The spring issues as drips and seeps from a rock overhang ("rain cave"), about 20 m above the river. There is no main orifice. The moist soil resulting from the spring discharge supports thick vegetation that extends down to the river. Samples were collected from the seeps with the highest discharge rate.

The spring issues from Miocene Patsy Mine volcanics (undifferentiated) near several northwest trending right lateral strike-slip faults. These faults offset low-angle faults, which produce the spring flow. A north-south trending high angle fault is located 0.5 km to the east of the spring.

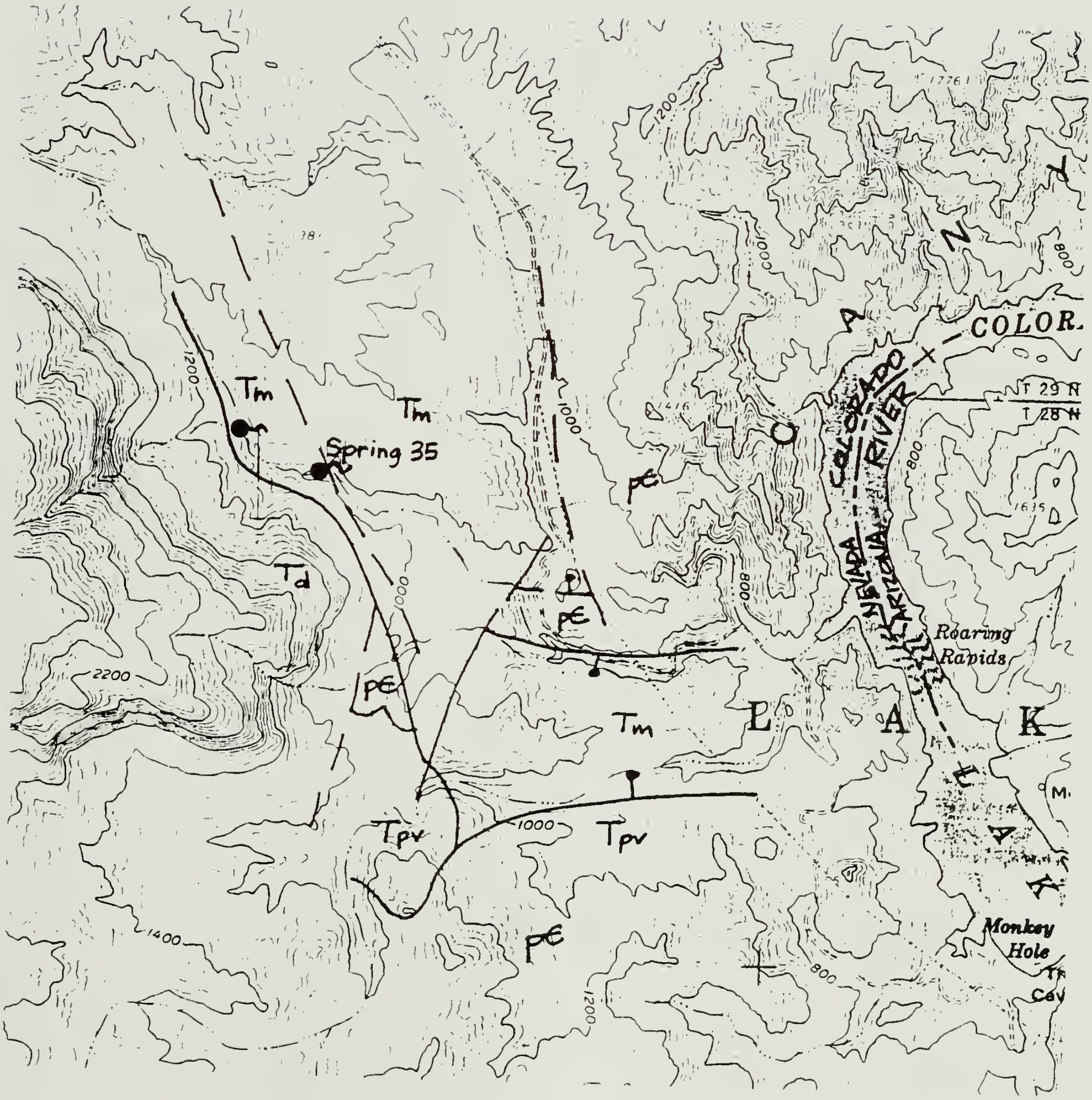


Spring 35 – Latos Pool

Topographic base: 7.5' Willow Beach Quadrangle
Geology references: Anderson (1978)

Latos Pool is located in Burro Wash on the eastern slope of the Eldorado Mountains, and about 1.6 km west of the Colorado River. Three pools fill a narrow portion of the wash, where the channel cuts through consolidated conglomerate. The two lower pools are connected and are both about 4 m long, 2 m wide, and over 1.5 m deep. The lowest pool is almost completely filled with reeds. The upper pool is smaller and is located about 15 m upstream. Another seepage area is located about 200 m upstream in a drainage extending from the southwest. This seep supports a thick stand of mesquite and grass. A third seepage area is located on a bench above the wash channel and about 100 m south of the pools. This seep also supports a thick stand of mesquite and grasses. Samples were collected from surface flow in the channel, below a seep area in the ravine walls and about 50 m below the pools. At the time of our visit (5-6-97), surface flow was discontinuous for about 100 m below the pools. However, evidence of recent surface flow (dried algae and salt deposits) extended from where the power line road crosses Burro Wash all the way upstream past the three pools.

Latos Pool is located on a fault trending N 15° W within a consolidated conglomerate of the Tertiary Muddy Creek Formation.



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ARIZONA
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T 28 N
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Rapids
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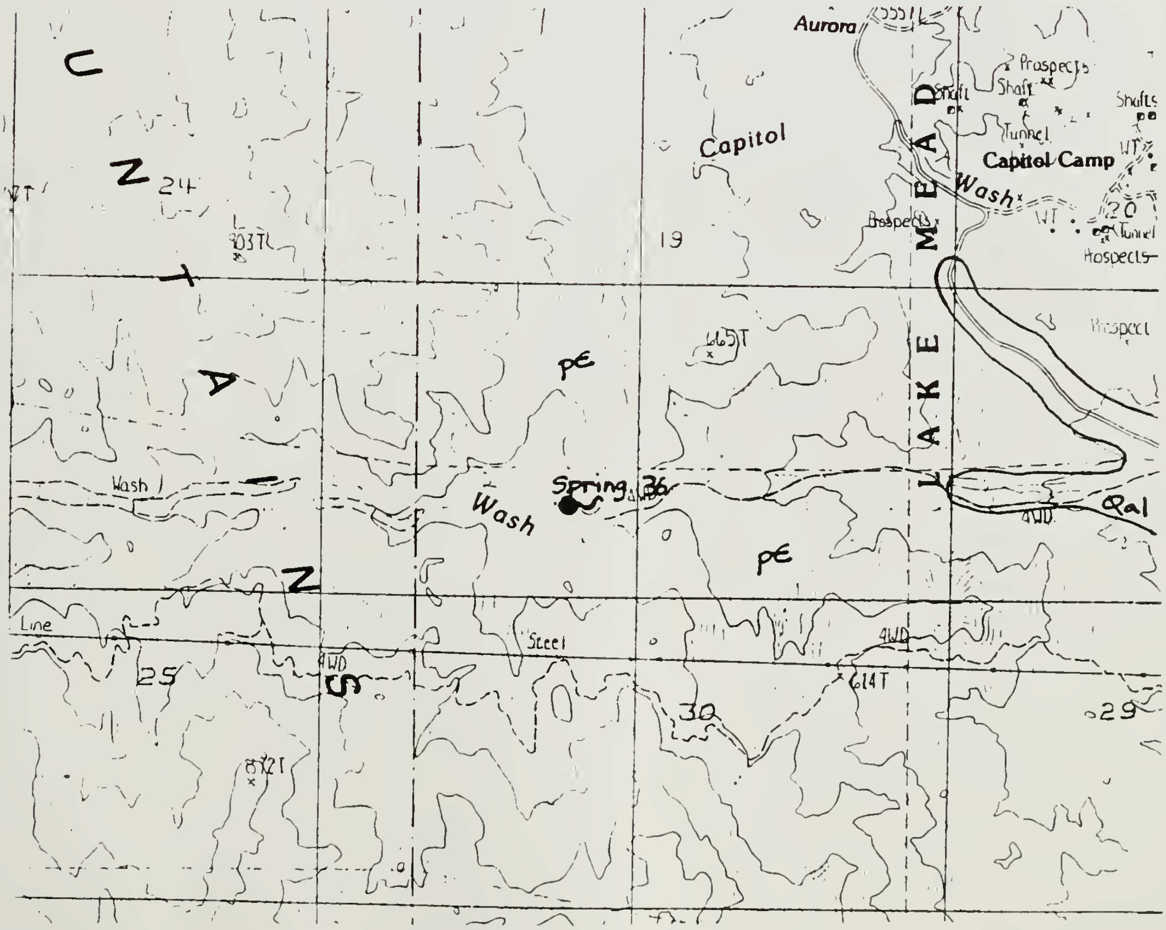
Spring 36 – Unnamed Spring in Aztec Wash

Topographic base: 7.5' Nelson Quadrangle
Geology references: Longwell (1963)

Aztec Wash extends out of the Eldorado Mountains toward Lake Mohave, south of Nelson, Nevada. The spring issues as seeps from alluvial deposits and directly from fractures in the granite walls of the channel, and flows discontinuously at the surface for about 150 m. The highest point of discharge was about 100 m upstream from a small concrete dam in the wash channel at our visit of 2-5-96. A pipe extends from the dam about 2 km down Aztec wash, although its poor condition prevents it from conveying any water. In addition, the channel behind the dam is filled with alluvial deposits. At the time of our second visit (2-11-97) the highest point of discharge was downstream about 20 m, and the discharge rate was lower. Grass and reeds line the banks of the ravine where a soil layer has developed; other areas consist of exposed granite with no soil or vegetation. Vegetation is sparse above the spring, suggesting that groundwater is deeper below ground surface.

The spring issues from steeply-dipping fractures in the Precambrian Rapakivi granite.



APPENDIX C

ISOTOPIC DATA FOR SELECTED SOUTHERN NEVADA GROUNDWATERS

Table C-1. Isotope Composition of Selected Southern Nevada Groundwaters. Values shown are averages if multiple samples are available. Number in parentheses is number of samples, if greater than one.

Site Name	Latitude (d m s)	Longitude (d m s)	Altitude of Land Surface (m AMSL)	δD (SMOW, ‰)	$\delta^{18}O$ (SMOW, ‰)	$\delta^{13}C$ (PDB, ‰)	PMC (uncorrected)	^{3}H (pCi/L)	Source*
McCullough Range, Eldorado Mountains, Highland Range, New York Mountains									
Crescent Spring	35 28 43	115 10 47	1292	-73.0	-9.4	—	—	—	c
Ora Hana Spring	35 37 25	115 04 07	1170	-72.0	-8.4	—	—	—	c
McClanahan Spring	35 41 42	115 11 05	902	-67.0	-7.2	-7.0	68.1	—	c
Rand Spring	35 42 03	114 51 20	1140	-78.0	-9.5	—	—	24.0	b
Bridge Spring	35 43 36	114 49 06	1032	-77.0	-9.2	—	—	19.0	b
Mormon Mountains									
Huckberry Spring	36 55 04	114 26 16	1580	-87	-12.3	—	—	—	c
Horse Spring	36 56 29	114 26 47	1750	-89	-12.7	—	—	—	c
Davies Spring	36 57 56	114 30 07	1825	-89	-12.5	—	—	—	c
East Mormon Mountains									
Peach Spring	36 57 16	114 17 23	950	-76.5	-10.4	—	—	—	c
Gourd Spring	36 57 31	114 17 30	950	-76.5	–10.6	—	—	—	c
Central Spring Mountains									
Trout Spring	36 13 22	115 40 59	2360	-97.7(19)	-13.6(22)	-8.1(5)	90.8(1)	257(3)	c
Cold Creek Spring	36 24 05	115 44 20	1930	-100.1(16)	-13.8(18)	-9.6(5)	76.0(4)	92(4)	c
Southern Spring Mountains									
Bird Spring	35 53 20	115 22 12	1326	-88.0	-11.7	-7.8	67.5	—	c
Sandstone Spring #1	36 03 47	115 28 09	1207	-89.0	-12.2	-10.6(2)	49.8(2)	<15(1)	c
BLM Visitors Center Well	36 07 44	115 26 03	1152	-89.0	-12.25	-9.3	46.0	9.0	c
Red Spring	36 08 40	115 25 10	1116	-89.0	-12.25	-10.5(2)	62.4(2)	3.0	c
Willow Spring	36 09 41	115 29 51	1402	-90.5	-12.3	—	—	—	c
White Rock Spring	36 10 27	115 28 43	1469	-91.0	-12.5	-12.0	—	<2.0	c
Castillo Well	35 50 02	115 26 09	1140	-94.0	-12.5	-9.3	39.4	—	c
Sheep Range									
Wiregrass Spring	36 38 00	115 12 29	—	-94.3(9)	-12.8(9)	-10.2	96.8	89.6	c
Moorman Well Spring	36 38 38	115 05 52	1963	-91.8	-12.7	-9.9	—	—	c
Cow Camp Spring	36 35 01	115 18 26	—	-92.0	-12.6	—	—	—	c
Lamb Spring	36 56 42	115 06 21	1700	-92.5	-13.15	—	—	—	c
Sawmill Spring	36 40 50	115 10 34	—	-92.0	-12.85	—	—	—	c
Sheep Spring	36 53 42	115 06 53	—	-96.0	-13.35	—	—	—	c
Meadow Valley Wash Flow System									
Wells and Springs	—	—	—	-87.3(14)	-11.8(13)	—	—	—	c
Lower White River Flow System									
Hiko Spring	37 35 54	115 12 49	—	-109.0	-13.8	-5.4	—	<10	e
Crystal Spring	37 31 58	115 13 50	—	-108(d)	-14.3(d)	-5.3	6.2	<10	e
Ash Spring	37 27 49	115 11 34	1102	-108.0	-14.1	-6.7	6.3	0.0	c
Big Muddy Spring	36 43 20	114 42 48	542	-97.8(3)	-12.9(3)	-6.0	6.7	<1.0	c
M–8 Spring	36 43 15	114 43 39	—	-99.0	-12.75	—	—	—	c
M–9 Spring	36 43 33	114 43 38	—	-96.5	-12.45	—	—	—	c

Table C-1. Isotope Composition of Selected Southern Nevada Groundwaters. Values shown are averages if multiple samples are available. Number in parentheses is number of samples, if greater than one (Continued).

Site Name	Latitude (d m s)	Longitude (d m s)	Altitude of Land Surface (m AMSL)	δD (SMOW, ‰)	$\delta^{18}O$ (SMOW, ‰)	$\delta^{13}C$ (PDB, ‰)	PMC (uncorrected)	^{3}H (pCi/L)	Source[1]
Lower White River Flow System Continued									
Pederson's Warm Spring	36 42 36	114 42 54	555	-97.0	-12.75	—	—	—	c
Iverson's Spring	36 42 37	114 42 43	—	-97.0	—	—	—	—	c
CE–VF–2 Well	36 52 30	114 56 44	752	-101.0(2)	-13.0(2)	-6.1	7.0	<1.0	c
CE–DT–6 Well	36 46 04	114 47 13	693	-97.0	-12.95	-8.0	8.4	1.8	c
CSV–2 Well	36 46 50	114 43 20	666	-98.0	-12.85	-5.5	8.4	4.0	c
Dry Lake Valley Well	36 27 18	114 50 38	638	-97.5	-13.3	-4.2	3.0	7.0	c
GP Apex Well	36 20 28	114 55 36	753	-98.0	-13.45	-5.5	2.7	<.3	c
CE–DT–4 Well	36 47 44	114 53 32	662	-101.0	-13.0	—	7.6	<2.0(1)	c
CE–DT–5 Well	36 47 44	114 53 32	661	-101.0	-13.0	—	7.6	<2.0(1)	c
Genstar Well	36 23 29	114 54 14	661	-97.0	-13.05	-4.9	1.5	<1.0	c
South Hidden Valley Well	36 33 08	114 55 30	807	-90.5	-11.2	—	—	—	c
CSV–3 Well	36 41 27	114 55 30	736	-75.0	-10.3	—	—	—	c
Weiser Wash Flow System									
EH–3 Well (Tmc)	36 41 32	114 31 32	530	-90.7(3)	-11.9(3)	—	—	—	d
EH–7 Well (Tmc)	36 40 14	114 31 53	512	-91.0	-12.3	—	—	—	d
EH–3 Well (below Tmc)	36 41 32	114 31 32	530	-92.0	-12.9	—	—	—	d
EH–7 Well (below Tmc)	36 40 14	114 31 53	512	-93.0	-12.8	—	—	—	d
Eldorado Valley									
Eldorado Substation Well	35 48 13	115 00 14	550	-96.0	-12.0	-7.8	7.75	<10	b
Colorado River									
Below Hoover Dam	36 00 35	114 44 40	200	-102.0	-12.7	-5.7	—	51.0	f
Valley of Fire									
Valley of Fire Well	36 25 21	114 32 52	683	-82.0	-10.6	-8.5	18.7	—	c
Northeast Las Vegas Valley									
Nellis AFB Well #13	36 12 44	115 03 00	552	-98.0	-13.8	-8.0	—	—	c
Lake Mead Base Well #3	36 14 21	115 00 16	568	-101.5	-13.8	-5.3	5.6	<.3	c
Nellis AFB #4	36 14 56	115 00 15	585	-95.0	-13.2	-6.3	21.0	—	c
Southwest Las Vegas Valley									
Sky Harbor Airport	35 58 16	115 08 50	—	-95.0	-13.1	-6.8	—	—	c
Showboat Country Club #2	36 02 51	115 04 48	—	-97.0	-13.3	—	—	—	c
Jean Prison Well	35 47 18	115 20 43	—	-95.0	-12.1	-7.6	2.4	—	c
Sunset Park Well	36 03 49	115 05 51	—	-94.0	-12.7	–6.7	4.0	—	c

[1] Sources of data:
a Thomas *et al.*, 1991
b SNWA, unpublished data
c Thomas, *et al.*, 1997
d DRI, unpublished data
e Hershey and Mizell, 1995
f This study